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Member of the Russian Academy of Sciences,
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Address for correspondence: Staromonetnyi per. 22, Moscow, 109180 Russia
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Belts of Arid and Humid Sedimentation during the Coniacian–Maastrichtian Ages of the Late Cretaceous

M. A. Zharkov

Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Abstract—The existence of six climatic belts—northern circumpolar humid coal, northern mid-latitude coal–bauxite–kaolinite, northern arid tropical–subtropical evaporite, equatorial humid coal–bauxite–kaolinite, southern arid tropical–subtropical evaporite, and southern humid coal–bauxite–kaolinite—are substantiated. Regular secular variations in latitudinal position of arid and humid belts related to global climate warming and general humidization at the end of the Late Cretaceous epoch are outlined.

Key words: *Coniacian, Santonian, Campanian, Maastrichtian, evaporite and coal-bearing basins, kaolinite- and bauxite-formation areas, arid and humid belts.*

INTRODUCTION

The paper continues a series of publications dedicated to paleogeography, deposition environments, and climatic zonality of the warm biosphere of the Cretaceous Period (Chumakov, 1995; Chumakov *et al.*, 1995; Filatova, 1995; Tsekhovskii *et al.*, 1995; Zharkov *et al.*, 1995, 1998a; Zharkov *et al.*, 1998b). It focuses on distinguishing the latitudinal belts of arid and humid sedimentation (Figs. 1–4) during the Coniacian, Santonian, Campanian, and Maastrichtian ages of the Late Cretaceous. Earlier, works of this kind were performed for older Cretaceous ages (Zharkov *et al.*, 1995, 1998a). To establish peculiar features of the spatial distribution of latitudinal climatic belts for every age of the second half of the Cretaceous, the schematic maps with belts of arid and humid sedimentation were compiled. They display the position of coal-bearing and evaporite basins, provinces and basins of redbed arid and gray-colored humid sedimentation, as well as areas and provinces of kaolinite and/or bauxite formation. Lithologic–paleogeographic maps for the Cretaceous age (Zharkov *et al.*, 1997b) served as a basis for the maps presented in this work. When compiling maps, published data, the majority of which were cited in previous papers (Zharkov *et al.*, 1995, 1998a, 1998b), were used. In addition, data from the following publications were taken into account: (Du Toit, 1954; Butov *et al.*, 1962; Shumilova, 1963; Houthon, 1966; Martinis and Visintin, 1966; Barnes, 1967; Averitt, 1969; Dobrovol'skii, 1969; Matveev, 1969, 1974; Blant, 1973; *Poverkhnosti* ..., 1974; Biske, 1975; *Cretaceous Facies*, 1977; Bowie *et al.*, 1978; *Kory* ..., 1979; Monakhov *et al.*, 1981; Dunning *et al.*, 1982; Franke *et al.*, 1982; Novikov, 1983; Dingle *et al.*, 1983; Cahen *et al.*, 1984; Patterson and Murray, 1984; Varnavskii, 1985; Pistsov and Muratova, 1987; Roehler, 1987; Skoblo and Lyamina, 1987; Vakhrameev and Lebedev, 1987; Liu Qun *et al.*, 1987; Vereshchagin *et al.*, 1987; Maccellari, 1988; Rodin, 1988; *Boksitonosnost'* ..., 1988; Smith, 1989; Bardossy and Aleva, 1990; Bardossy and Dercourt, 1990; Bussert *et al.*, 1990; Germann *et al.*, 1990; Luger *et al.*, 1990; Wycisk, 1990; Rodin and Filatova, 1991; Camoin *et al.*, 1993a; Camoin *et al.*, 1993b; Golonka *et al.*, 1995; Robinson and Kirschbaum, 1995; Rouchy *et al.*, 1993; Sitian *et al.*, 1995).

Indeed, it is impossible to take fully into account all the data published. Therefore, some data on certain regions or countries may be missing. Despite this, the data used in this work allowed the belts and areas of arid and humid sedimentation to be established and outlined for all the ages of the second half of the Late Cretaceous. It should be noted that latitudinal zoning is mainly recognized based on climatic indicators of arid and humid sedimentation settings on continents (Strakhov, 1963). Information on climatic conditions in oceans and shelf and epicontinental seas is used as a supplementary indicator.

Equatorial Humid Coal–Bauxite–Kaolinite Belt

The existence of the equatorial humid coal–bauxite–kaolinite belt, which is clearly manifested in the territory of both the African and South American continent, is one of the most remarkable features of the climatic zonality during the second half of the Late Cretaceous. Within the African continent, the belt comprises two large provinces of the kaolinite and/or bauxite formation: (1) the Central African province extending from Cameroon in the west to Sudan in the east and (2) the Leo-Man province occupying the territory of modern Guinea, southern part of Mali, and northern half of the Ghana territory in southwestern Central

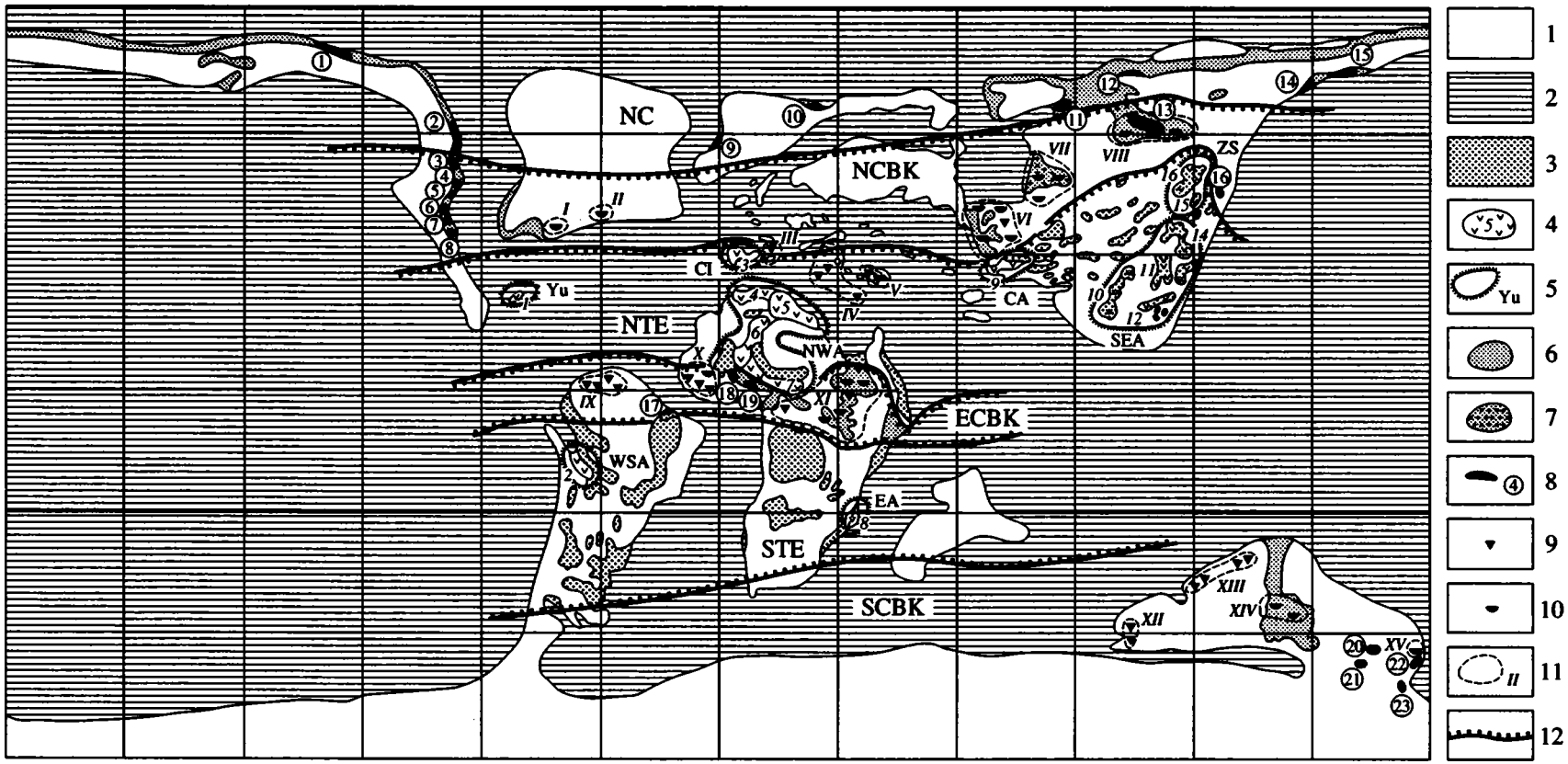

Fig. 1. Belts and areas of arid and humid sedimentation during the Coniacian Age of the Late Cretaceous.

(1) dry land; (2) oceans, shelf and epicontinental seas; (3) areas and basins of redbed arid sedimentation; (4) evaporite-bearing basins: (*1*) Yucatan, (2) Altiplano (Potosi, Chita, and others), (*3*) Central Iberian, (*4*) Tindouf, (*5*) Algerian–Tunisian, (*6*) Tanezzrouft, (*7*) Mali–Nigerian, (*8*) Uanetze, (*9*) Central Asian, (*10*) Lanpang-Simao, Dien-Zsu (*11*) Jianghan, (*12*) Hefu, Ditay, Ganzan, and others, (*13*) Nanchang, (*14*) Subei, Northern Chinese, (*15*) Songnlao, (*16*) Zeya–Bureya; (5) evaporite provinces and their indices: (Yu) Yucatan, (WSA) Western South American, (NWA) Northwest African, (CI) Central Iberian, (EA) East African, (CA) Central Asian,(ZS) Zeya–Songlao, (SEA) Southeast Asian; (6) basins and areas of gray-colored terrigenous humid sedimentation; (7) basins and areas of variegated and redbed humid sedimentation; (8) coal-bearing basins and their numbers: (*1*) Colville, (2) Rocky Mountains and Foothills, (*3*) Northern Central Montana, (*4*) Jackson Hole, (*5*) Hems Fork, (*6*) Green River, Wind River, (*7*) Uinta, (*8*) San Juan, (*9*) Western Greenland, (*10*) Northern Greenland, (*11*) Eastern Khatanga, (*12*) Novosibirskie Islands, (*13*) Lena, (*14*) Penzhina, (*15*) Anadyr', (*16*) Partizansk, Razdol'naya, (*17*) Barreirinhas, (*18*) Sokoto, (*19*) Gao, (*20*) Gippsland, Strzelecki, (*21*) Bass, (*22*) Greymouth, Paparoa, Brunner, and others, (*23*) Kaintangata, Ohai, and others; (9) areas of bauxite formation; (10) areas of kaolinite formation; (11) areas of kaolinite and/or bauxite formation and their numbers: (*I*) Gulf Coastal Plains, (*II*) Northeastern United States, (*III*) Langedoc–Provence, (*IV*) Tatra–Dinarids–Taurus, (*V*) Pelagonian, (*VI*) Northern Kazakhstan, (*VII*) Central Siberian, (*VIII*) Vilyui–Verkhoyan'e, (*IX*) Guiana, (*X*) Leo-Man, (*XI*) Central African, (*XII*) West Australian, (*XIII*) North Australian, (*XIV*) Eromanga, (*XV*) New Zealand; (12) belt boundaries. Belt indices: (NC) northern circumpolar humid coal belt, (NCBK) northern mid-latitude humid coal–bauxite–kaolinite belt, (NTE) northern tropical–subtropical arid evaporite belt, (ECBK) equatorial humid coal–bauxite–kaolinite belt, (STE) southern tropical–subtropical arid evaporite belt, (SCBK) southern humid coal–bauxite–kaolinite belt.

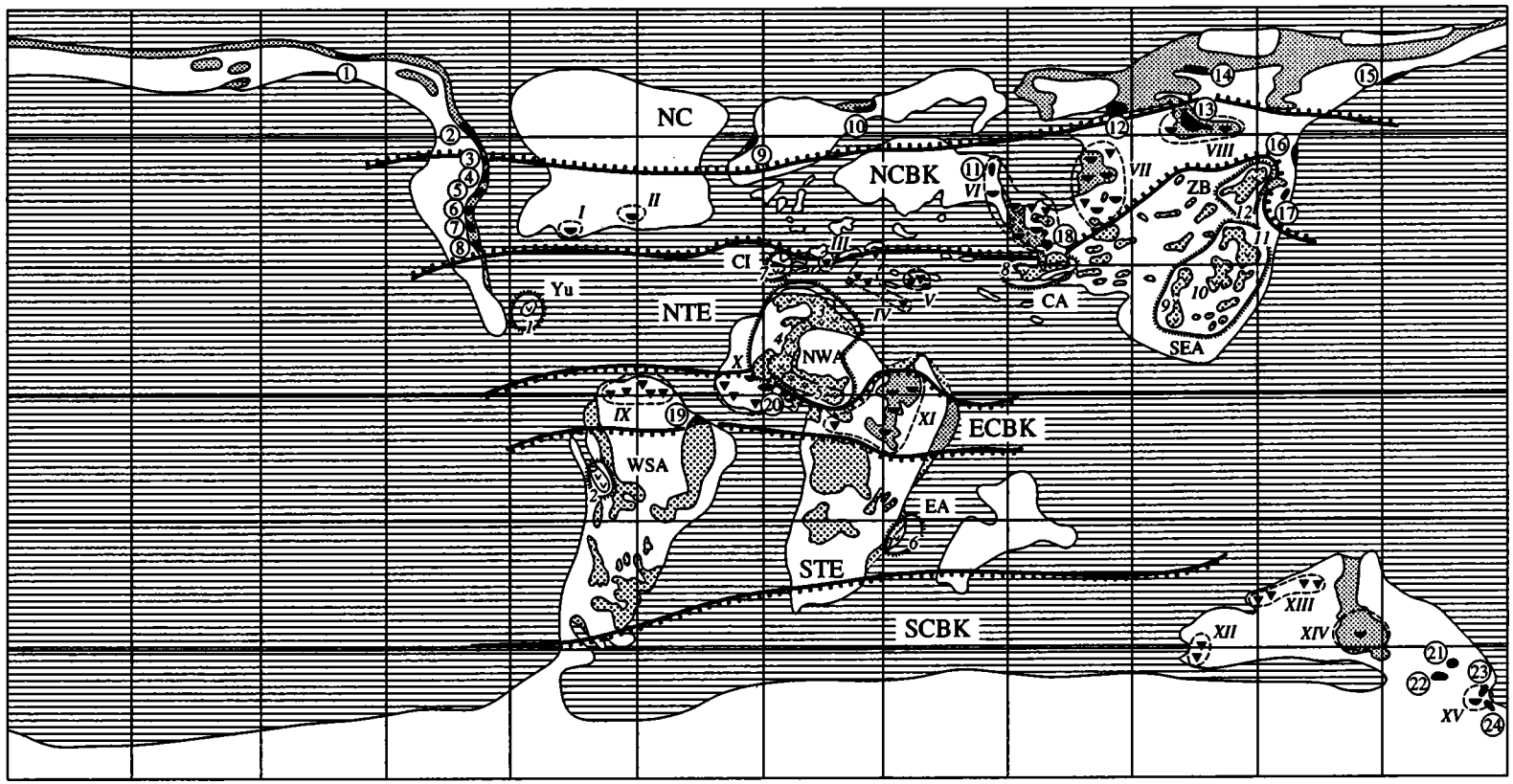

Fig. 2. Belts and areas of arid and humid sedimentation during the Santonian Age of the Late Cretaceous. For legend see Fig. 1.

Evaporite-bearing basins: (*1*) Yucatan, (2) Altiplano, (*3*) Algerian–Tunisian, Tindouf, (*4*) Tanezzrouft, (*5*) Mali–Nigerian, (*6*) Uanetze, (*7*) Central Iberian, (*8*) Central Asian, (*9*) Lanpang-Simao, Dien-Zsu, (*10*) Jianghan, (*11*) Subei, (*12*) Zeya–Bureya. Evaporite provinces: (Yu) Yucatan, (WSA) Western South American, (NWA) Northwestern African, (EA) East African, (CI) Central Iberian, (CA) Central Asian,(ZS) Zeya–Bureya, (SEA) Southeast Asian. Coal-bearing basins: (1) Collwill, (2) Rocky Mountains and Foothills, (3) Northern Central Montana, (4) Jackson Hole, (5) Hems Fork, (6) Green River, Wind River, (7) Uinta, (8) San Juan, (9) Western Greenland (Disco Islands), (10) Northern Greenland (Wandel Sea), (11) North Uralian, (12) Eastern Khatanga, (13) Vilyui, Lena, (14) Novosibirskie Islands, (15) Anadyr', (16) Western Sikhote Alin, (17) Partizansk, Razdol'naya, (18) Eastern Turgai, (19) Barreirinhas, (20) Gao, (21) Gippsland, Strzelecki, (22) Bass, (23) Greymouth, Paparoa, and others, (24) Kaintangata, Ohai, and others. Areas of kaolinite and/or bauxite formation: (*I*) Gulf Coastal Plains, (*II*) Northeastern United States, (*III*) Langedoc–Provence, (*IV*) Tatra–Dinarids–Taurus, (*V*) Pelagonian, (*VI*) Kazakhstan–Uralian, (*VII*) Central Siberian, (*VIII*) Vilyui–Verkhoyan'e, (*IX*) Guiana, (*X*) Leo-Man, (*XI*) Central African, (*XII*) West Australian, (*XIII*) North Australian, (*XIV*) Eromanga, (*XV*) New Zealand.

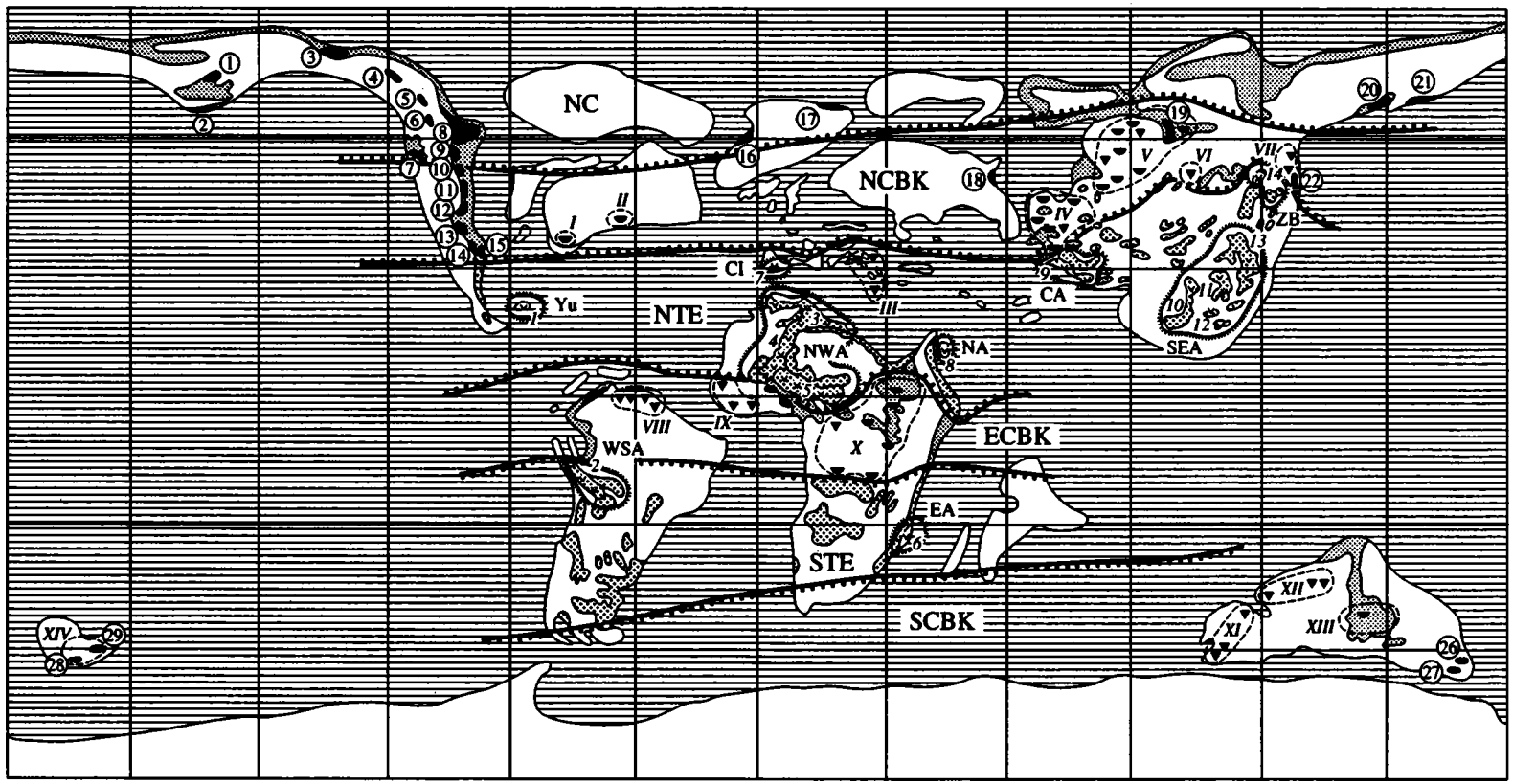

Fig. 3. Belts and areas of arid and humid sedimentation during the Campanian Age of the Late Cretaceous. For legend see Fig. 1.

Evaporite-bearing basins: (*1*) Yucatan, (*2*) Altiplano (Potosi, Chita, and others), (*3*) Algerian–Tunisian, Tindouf, Libyan, (*4*) Tanezzrouft, (*5*) Mali–Nigerian, (*6*) Mozambique–Tanzania, (*7*) Central Iberian, (*8*) Northern Arabian, (*9*) Central Asian (Fergana and others), (*10*) Lanpang-Simao, (*11*) Jianghan, (*12*) Hefu, Ditay, and others, (*13*) Subei, (*14*) Zeya–Bureya. Evaporite provinces: (Yu) Yucatan, (WSA) Western South American, (NWA) Northwest African, (EA) East African, (Cl) Central Iberian, (NA) North Arabian, (CA) Central Asian, (ZS) Zeya–Bureya, (SEA) Southeast Asian. Coal-bearing basins: (1) Rampart, (2) Chingnik, (3) Collville, (4) Peel, (5) Fraser, (6) Sastut, (7) Nanaimo, (8) West Canadian (Alberta, Rocky Mountains and Foothills, and others), (9) Northern Central Montana, (10) Bighorn, (11) Green River, Wind River, (12) Uinta, (13) San Juan, (14) Ojinati, (15) Sabinas, (16) Western Greenland (Disco Island, Svartenhuk), (17) Northern Greenland (Wandel Sea), (18) North Uralian, (19) Vilyui–Verkhoyansk, (20) Penzhina, (21) Anadyr', (22) Partizansk, Razdol'naya, (23) Middle Magdalena, Western Maracaibo, (24) Gao, (25) Benin, (26) Gippsland, Strzelecki, (27) Bass, (28) Greymouth, Paparoa, and others, (29) Kaintangata, Ohai, and others. Areas of kaolinite and/or bauxite formation: (*I*) Gulf Coastal Plains, (*II*) Northeastern United States, (*III*) Tatra–Dinaridss–Taurus, (*IV*) Northern Kazakhstan–Siberian, (*V*) East Siberian, (*VI*) Vitim–Patom, (*VII*) Sikhote-Alin, (*VIII*) Guiana, (*IX*) Leo-Man, (*X*) Central African, (*XI*) West Australian, (*XII*) North Australian, (*XIII*) Eromanga, (*XIV*) New Zealand.

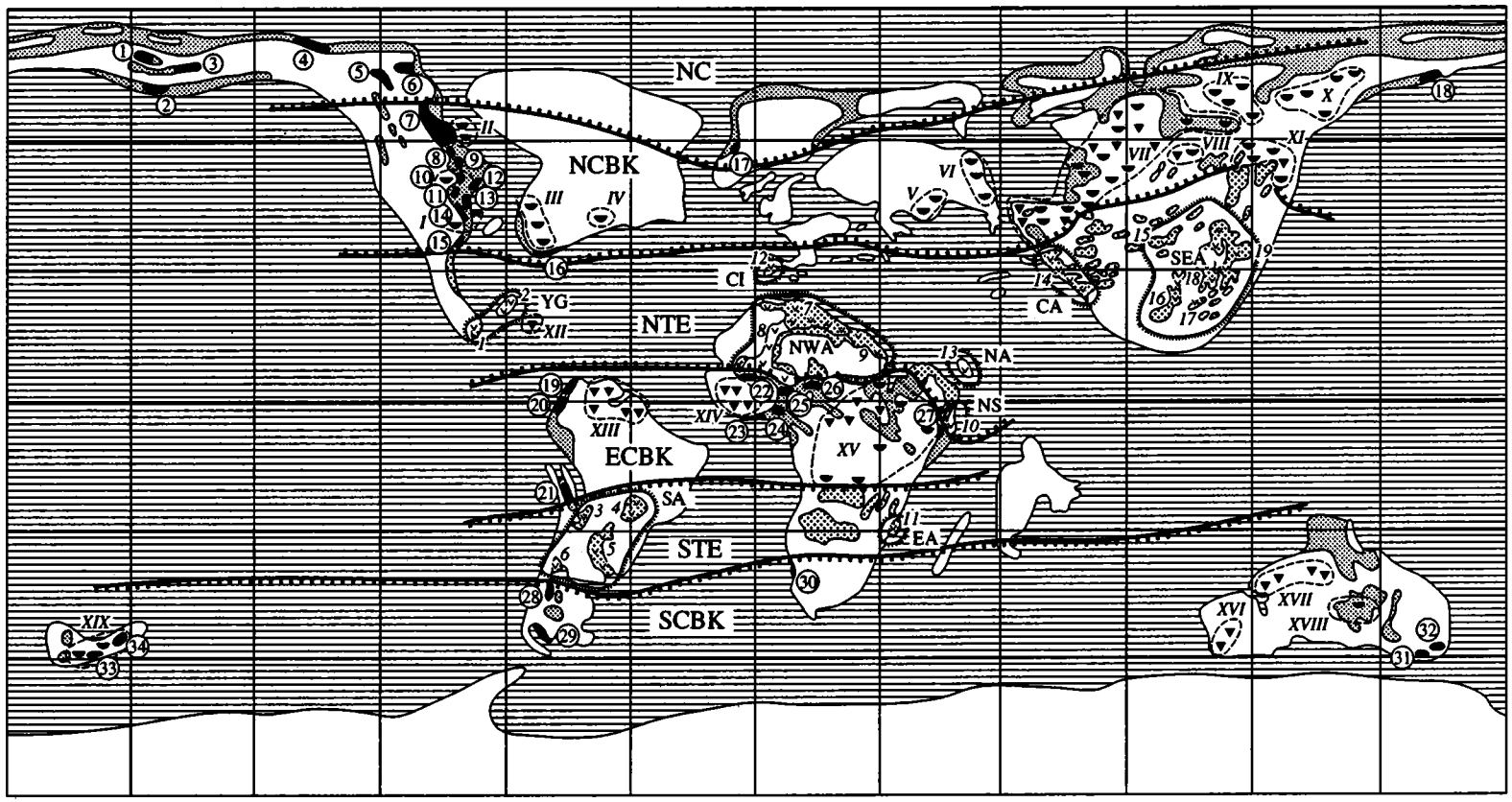

Fig. 4. Belts and areas of arid and humid sedimentation during the Maastrichtian Age of the Late Cretaceous. For legend see Fig. 1.

Evaporite-bearing basins: (*1*) Honduras, (*2*) Yucatan, (*3*) Northwestern Subandean, (*4*) Parana, (*5*) Chaco, Salado, (*6*) Neuqen, (*7*) Algerian–Tunisian, Tindouf, (*8*) Tanezzrouft, Sokoto, (*9*) West Libyan, Sirte, (*10*) Al-Mado, (*11*) Mozambique–Tanzania, (*12*) Central Iberian, (*13*) Tayarat, (*14*) Central Asian (Fergana and others), (*15*) Chaoshui-Hetao, (*16*) Lanpang-Simao, (*17*) Hefu, Ditay, (*18*) Jianghan, (*19*) Subei and others. Evaporite provinces: (YG) Yucatan–Honduras, (SA) South American, (NWA) Northwest African, (NS) Northern Somali, (EA) East African, (CI) Central Iberian, (NA) North Arabian, (CA) Central Asian, (SEA) Southeast Asian. Coal-bearing basins: (1) Stuart, (2) Chingina, (3) Rampart, Doll, Nenana, (4) Collville, (5) Peel, (6) Western Mackenzie, (7) West Canadian (Alberta and others), (8) Northern Central Montana, (9) Bighorn, Powder River, (10) Green River, Wind River, (11) Uinta, (12) Denver, (13) Raton, (14) San Juan, (15) Sabinas, (16) Ripley McNairy, (17) Western Greenland, (18) Anadyr', (19) Barko, Carare, (20) Bogota, Boyaca, (21) Chicamos, Hunin, (22) Gao, (23) West Nigerian (Dahomeyan), (24) South Nigerian, (25) Benua, (26) Chad–Nigerian, (27) Hed-Hed, (28) Chubut, (29) Northern Magallanes, (30) Orange, (31) Bass, (32) Gippsland, Strzelecki, (33) Greymouth, Paparoa, and others, (34) Kaintangata, Ohai, and others. Areas of kaolinite and/or bauxite formation: (*I*) San Juan–Raton–Green River, (*II*) Williston, (*III*) Mississippi, (*IV*) Northeastern United States, (*V*) North Moscowian, (*VI*) Uralian, (*VII*) Kazakhstan–Siberian, (*VIII*) Vitim–Patom, (*IX*) Yana–Indigirka, (*X*) Kolyma–Okhotsk, (*XI*) Amur–Sikhote-Alin, (*XII*) Cuba, (*XIII*) Guiana, (*XIV*) Leo-Man, (*XV*) Central African, (*XVI*) West Australian, (*XVII*) North Australian, (*XVIII*) Eromanga, (*XIX*) New Zealand.

Africa. Humid environments favorable for coal accumulation also occurred intermittently in these regions. Such settings were confined, as a rule, to the coastal areas of the Trans-Saharan seaway. In the Coniacian, coal accumulation occurred in the Sokoto and Gao basins; in the latter, it continued till the Maastrichtian. The Benin coal-bearing basin was formed in the Campanian. The Maastrichtian was marked by substantial expansion of coal accumulation environments. Such conditions are recognized in the West (Dahomeyan) and South Nigerian, Benin, and Chad–Nigerian basins, as well as in the Hed-Hed basin of Somali in eastern Africa. In South America, the belt comprised the Guiana bauxite-bearing province and coal-bearing Barreirinhas basin, which existed during the Coniacian and Santonian ages, as well as the Middle Magdalena, Western Maracaibo, Barko, Carare, Bogota, Boyaca, Chikamos, Hunin, and other basins, where coal accumulation occurred in the Campanian and, mainly, Maastrichtian.

The northern boundary of the coal–bauxite–kaolinite belt is traced near the equator. It remained almost unchanged during the entire second half of the Late Cretaceous. The boundary is most distinct on the African continent, where it passes between closely spaced neighboring provinces and areas of coal, kaolinite, and/or bauxite accumulation, on the one hand, and areas of evaporitic redbed arid sedimentation, on the other. Near the eastern coast of Africa, the boundary sharply turns to the south. There, its position for the Maastrichtian Age is established between the Hed-Hed coal-bearing basin and Al-Mado saliferous basin in northern Somali. A similar southward shift of the northern boundary is also assumed for the Coniacian, Campanian, and Maastrichtian ages, as it is evident from the development of redbed arid deposits in the coastal areas of the Arabian basin. The western segment of the northern boundary of the equatorial belt is arbitrarily outlined near the northern margin of the South American continent.

The location of the southern boundary of the equatorial belt is also rather reliably established between neighboring humid and arid provinces. This boundary gradually migrated with time southward, because of expansion of kaolinite- and bauxite-formation environments in the Central African province recorded in Uganda, Kenya, Congo, and northern Zambia, and also due to growth of coal accumulation settings in western areas of South America, particularly in Peru (Chikamos, Hunin, and other coal-bearing basins). Whereas, in the Santonian, the southern boundary was located near 10°–20°S; in the Maastrichtian, it shifted slightly south of 30°S. As a result, a substantial extension of the equatorial humid belt occurred after the Santonian, and it was transformed from a relatively narrow (no more than 2500–2700 km) into a sufficiently wide (about 5000 km) and distinctly traceable zone.

The Northern Arid Tropical–Subtropical Evaporite Belt

The northern arid belt retained its position and almost unchanged boundaries during the entire stage of the Late Cretaceous history under consideration. It comprised southern areas of Laurasia, northern Africa, and almost all the Tethyan realm. One of the largest regions of arid sedimentation occupied the northern and northwestern areas of the African continent. This region was almost permanently characterized by climatic conditions favorable for evaporitic sedimentation, both in coastal basins connected with the Tethys (Algerian–Tunisian, Tindouf, Western Libyan, Sirte, and other basins) and within the Trans-Saharan seaway (Mali–Nigerian, Sokoto, and Tanezzrouft basins). Over the prolonged periods, spacious areas in this region hosted settings of redbed coastal and lacustrine sedimentation frequently displaying the desert-type features. In the Campanian and Maastrichtian, evaporitic sedimentation recommenced in the North Arabian and northern Somali provinces (Tayarat, Hed-Hed, and other basins).

The second large region of arid environments continued to exist in the southeastern Laurasian continent. This region inherited two provinces of evaporitic sedimentation in southeastern and central Asia. They included numerous continental sulfate-bearing basins, the largest among which were Central Asian, Lanpang-Simao, Dien-Zsu, Jianghan, Hefu, Ditay, Ganzan, Subei, Chaoshui-Hetao, and other basins. Beginning from the Coniacian, the new Zeya–Songliao evaporitic province appeared in the eastern Laurasian continent, where the continental lacustrine to alluvial redbed and sulfate sedimentation is registered in the Shaulao basin in northeastern China and Zeya–Bureya basin in the central Amur region. Arid environments of the redbed terrigenous deposition were also characteristic of many areas of Mongolia, the Transbaikal region, and Olekma–Vitim Highland. In general, it can be noted that the second half of the Late Cretaceous was a time period of substantial expansion of arid sedimentation settings in eastern Laurasia as compared with the Middle Cretaceous, and they migrated northward in many interior areas of the continent.

The third large region with intermittently existed arid sedimentation environments covered almost the entire Tethyan realm. There, favorable conditions for evaporite accumulation were developed in many pericratonic carbonate platforms. Evaporitic basins of this type are recorded in the Yucatan, Central Iberian, North Arabian, and some other regions. It should be noted that they were localized both in northern and southern areas of the Tethyan province. In this connection, it should be emphasized that carbonate platforms of the Cretaceous time, including the Late Cretaceous, were mostly confined to the Tethys in the northern hemisphere, i. e., mainly to northern tropical and subtropical zones, which is thought to be related to circumglobal warm western currents (Camion *et al.*, 1993a; Camion

et al., 1993b; Dercourt *et al.*, 1993; Masse *et al.*, 1993; Philip *et al.*, 1993). Spatial distribution of carbonate platforms is meridionally asymmetric: they are mainly located on eastern margins of continents (the southeastern periphery of North America, eastern areas of the African–Arabian craton) and along the southern periphery of interior areas of Laurasia. Their latitudinal position was stable within the belt between 18–35°N in the Caribbean, Mediterranean, and Eastern Tethys (Camion *et al.*, 1993). In fact, during the entire second half of the Late Cretaceous, settings of shallow water biogenic carbonate accumulation were characteristic of the northern arid belt. This allows the location of carbonate platforms to be used as an additional indicator for defining the belt boundaries. At the same time, it should be kept in mind that climatic conditions were not uniformly and permanently arid in the Tethyan realm. In some areas, they intermittently became warm and humid, which is evident from the existence of bauxite-accumulation settings in carbonate platforms of the Langedoc–Provence, Tatra–Dinarids–Taurus, Pelagonian, and Cuban (in the Maastrichtian) regions.

In most of the Tethyan areas, the northern boundary of the northern arid belt is drawn approximately along northern margins of carbonate platform distribution areas, where it is traced between 30°N and 35°N. In the southern part of North America, the boundary changed its position, gradually shifting from 30°N to 32°N. In the east, in the Asian segment of Laurasia, it first passes in the northwestern direction to coincide with 50°N in the Amur River lower courses and then sharply turns toward the south and southeast. The boundary location in this province is sufficiently clear everywhere: it runs between basins with evaporitic and redbed arid sedimentation, on the one hand, and closely spaced coal-bearing basins and areas of the kaolinite and bauxite formation, which are located within the northern humid coal–bauxite–kaolinite belt, on the other.

The Southern Arid Tropical–Subtropical Evaporite Belt

The southern arid belt is clearly manifested on the territory of both South American and African continents. In South America, it comprised many continental and coastal-marine evaporitic basins, whose number and spatial distribution substantially varied with time. The following basins can be mentioned among them: Potosi, Chita, Northwestern Subandean, Parana, Chaco, Salado, and Neuqen. In the Coniacian, Santonian, and Campanian, evaporitic basins were mainly confined to the northern areas of the belt, whereas the remaining territory represented the areas of the prevalent alluvial–lacustrine redbed and variegated terrigenous sedimentation. The Maastrichtian was marked by extension of evaporite accumulation environments almost over the entire territory of the considered belt up to its southern boundaries. In the African continent, the sole East African evaporitic province is established based on the presence of evaporites in the Uanetze and Mali–Nigerian basins. Many other interior and coastal areas of this continent (Congo, Okawango, Orange, probably, Kalahari, and others) mainly hosted the continental redbed sedimentation settings.

The southern boundary of the arid belt is arbitrarily outlined between 35°S and 40°S. In South America, it is unequivocally established between the Chubut coal-bearing and Neuqen evaporite-bearing basins only in the Maastrichtian. In response to the above-mentioned southward extension of the equatorial coal–bauxite–kaolinite belt, the southern arid belt gradually narrowed from the approximately 4500–4700 km wide in the Coniacian–Santonian to the 2500–3500 km wide in the Maastrichtian.

The Northern Mid-Latitude Humid Coal–Bauxite–Kaolinite Belt

During the entire second half of the Late Cretaceous, the northern mid-latitudes of Laurasia were occupied by humid settings of coal accumulation as well as of bauxite and kaolinite formation. This allows one to distinguish the northern humid coal–bauxite–kaolinite belt, which is well traced from western to eastern margins of the Laurasian continent.

On the territory of North America, the mid-latitude humid belt included, in the Coniacian to Campanian time, two kaolinitic provinces: the Gulf Coastal Plains and the northeastern areas of the United States. In the Maastrichtian, the kaolinite-formation environments spread over many peripheral areas of the Rocky Mountains and coastal plains of the Western Interior seaway. One of these kaolinite-accumulation provinces was located in the southern part of the belt near the San-Huan, Raton, and Green River basins, and the other one, in the north, within the Willystone basin. A number of coal-bearing basins were also localized in this territory during the second half of the Late Cretaceous. The largest of them were the Sabinas, Ojinaga, San Juan, Uinta, Raton, Wind River, Green River, Bighorn, Hems Fork, Jackson Hole, and Alberta basins. In the Campanian and Maastrichtian, the belt significantly widened in the north direction. Simultaneously, its boundary shifted from 55–57°N to 68–70°N.

In the Asian part of Laurasia, the territory of the northern mid-latitude humid belt was always characterized by the particularly significant development of kaolinite- and bauxite-formation settings. They changed configuration and dimensions and were gradually distributed over spacious regions. In the Coniacian, such environments were characteristic of three provinces: northern Kazakhstan, central Siberia, and Vilyui–Verkhoyan'e regions. Beginning in the Campanian, bauxite- and kaolinite-formation environments advanced to the most eastern areas of Laurasia. They became concentrated in the Kazakhstan–West Siberia, East Siberia, Vitim–Patom, and Sikhote Alin' regions. In the Maastrichtian, such conditions were characteris-

tic of not only eastern areas of the Ural–Baltia region, where two, North Moscowian and Uralian kaolinitic provinces are distinguished, but also of all eastern and northern regions of Asia from Kazakhstan on the west to the Far East and Chukchi region on the east. Five large provinces of kaolinite and bauxite formation are outlined: the Kazakhstanian–Siberian, Vitim–Patom, Yana–Indigirka, Kolyma–Okhotsk, and Amur–Sikhote Alin' regions.

In Asia, the coal-formation areas and a series of coal-bearing basins were reduced during the second half of the Late Cretaceous. Only two coal-bearing basins are established in the west of this region: the North Urals and East Turgai basins. In the east, coal accumulation continued till the Maastrichtian in the Lena, Vilyui, western Sikhote Alin', Partizansk, Razdol'naya, and some other basins. In the Maastrichtian, the mid-latitude humid belt advancing northward comprised coal-bearing basins of the Anadyr' area.

It should be noted that the northern mid-latitude humid belt is rather confidently established only in two distant regions: North America on the west and Asia on the east. Central regions of the belt lack any sedimentological evidence of a continental humid climate. These regions were covered by spacious epicontinental sea basins with specific sedimentation settings. The largest among them were the North Sea, Polish–Lithuanian, Central European, East European, Circum-Caspian, Turan, Afghanistan–Pakistan, West Siberian basins, and also the Western Interior basin in North America. Many of them hosted chalk, phosphate, glauconite, ferruginous, or siliceous sedimentation. The confinement of such basins to the mid-latitude humid belt was probably not incidental. Sedimentation in them was strongly influenced by humid climate and by influx of redeposited weathering products transported from the humid provenance.

In general, it may be noted that the northern mid-latitude humid belt is traceable with a sufficient confidence by using the distribution patterns of the continental settings of kaolinite, bauxite, and coal formation, as well as the marine settings of chalk, glauconite, phosphate, iron, and siliceous accumulation in epicontinental seas. As was mentioned, the southern boundary of the belt largely passed between 30°N and 35°N and, only in the Asian part of Laurasia, it shifted far northward almost to 50°N and then sharply turned to the south and southeast. The northern boundary permanently changed its position. In the Coniacian and Santonian, it was located between 50°N and 65°N. Beginning in the Campanian, the boundary progressively migrated to 70°N and 75°N, where it is recorded both in North America and Asia. As a result, the significant expansion of the northern mid-latitude humid belt occurred in the Maastrichtian, especially in the east, where kaolinite-formation settings spreaded over the entire northeastern part of Laurasia.

The Southern Humid Coal–Bauxite–Kaolinite Belt

The southern coal–bauxite–kaolinite belt is clearly defined only in Australia and New Zealand. There, it is established by joint or close occurrence of kaolinite-, bauxite, and coal-formation settings. These environments are distinctly established for both the Coniacian and Santonian ages, when Australian and New Zealand plates were coupled with Antarctica to form a single continent, and also for the Campanian and Maasrtichtian after break-up of the latter. The western and interior areas of Australia always hosted three large, West Australian, North Australian, and Eromanga basins of kaolinite and bauxite accumulation, coexisting with the Gippsland, Strzelecki, Bass, Greymouth, Paparoa, Brunner, Kaitangata, and Ohai coal-bearing basins in southeastern Australia and kaolinite basin in New Zealand. During the Coniacian–Santonian, these areas were located between 40°S and 75°–78°S, but, in the Campanian and Maastrichtian, they occupied, owing to the northward movement of Australia and New Zealand, latitudes of 40°–60°S.

Separate areas of the humid belt in the extreme south of Africa and in South America are recognizable only for the Maastrichtian Age. At that time, the humid belt in southern and southwestern Africa comprised the Orange lignite-bearing basin and wide areas of siliceous weathering crust (silcretes) in the Cape province. A great number of Maastrichtian coal-bearing basins are established in southern areas of South America, among which the North Magellanes and Chubut basins should be mentioned. Taking into consideration the latter, the northern boundary of the belt can be outlined. In the Maastrichtian, it was located near 40°–44°S. Similar or close position of this boundary was likely characteristic for older ages of the Late Cretaceous as well. The position of the belt southern boundary remains yet unknown. The belt probably included also a significant part of Antarctica, but there are no data supporting this assumption.

The Northern Circumpolar Humid Coal Belt

The northern humid belt is outlined based solely on the distribution of continental and maritime coal-formation settings in subpolar areas of Laurasia. They are established in many basins of northwestern America, Greenland, and Asia.

It is remarkable that distribution of coal-accumulation settings in western and eastern sectors of the northern coal belt was asymmetric during the entire second half of the Late Cretaceous. They always were concentrated in the territory of North America. A number and total area of coal-bearing basins progressively increased. For instance, during the Coniacian and Santonian ages, the coal accumulation was in progress in the Collville basin and in basins of the Rocky Mountains and Foothills and also of western and northern Greenland. In the Campanian, coal accumulated in the

Peel, Fraser, Sastut, Nanaimo, and Alberta basins, and during the Maastrichtian, in the Stuart, Chignina, Doll, Nenana, and Western Mackenzie basins. Simultaneously, coaliferous deposits in the Asian sector of the subpolar belt were localized in few basins of the eastern Khatanga, Novosibirskie Islands, Penzhina, and Anadyr' regions. In the Maastrichtian, all these basins were constituent areas of the southerly located humid coal–bauxite–kaolinite belt.

It should be noted that available materials allow the circumpolar coal belt to be outlined only in the northern hemisphere of the Earth while data on the southern hemisphere are extremely scarce.

CONCLUSION

(1) Principal features of latitudinal zoning in distribution of arid and humid sedimentation settings at the early stages of the Late Cretaceous (Zharkov *et al.*, 1995) were also characteristic of the second half of the Late Cretaceous. Six latitudinal belts of the last epoch with arid or humid climate can be distinguished: (1) northern circumpolar humid coal belt; (2) northern mid-latitude humid coal–bauxite–kaolinite belt; (3) northern arid tropical–subtropical evaporite belt; (4) equatorial humid coal–bauxite–kaolinite belt; (5) southern arid tropical–subtropical evaporite belt; and (6) southern humid coal–bauxite–kaolinite belt. These latitudinal belts are recognized based on distribution patterns of arid sedimentation settings with redbeds and evaporites, and of humid settings with coal, bauxite, and kaolinite studied in continents. Boundaries and strike of the northern arid evaporite belt are additionally substantiated by shallow-water accumulation settings of biogenic carbonates widely developed in the Tethyan carbonate platforms. The location and boundaries of the northern mid-latitude humid coal–bauxite–kaolinite belt are also justified by the development of epicontinental marine basins with chalk, glauconite, phosphate, iron, and siliceous sedimentation typical of them.

(2) The second half of the Late Cretaceous was marked by regular changes in latitudinal position of humid and arid belts related most probably to global climate fluctuations. These transformations first affected the latitudinal belts of humid sedimentation. For instance, in the Campanian and, mainly, in the Maastrichtian, the northern humid coal–bauxite–kaolinite belt significantly widened and advanced northward, whereas the equatorial humid coal–bauxite–kaolinite belt experienced a significant southward extension. Simultaneously, the width of the southern arid evaporite belt was universally reduced. All these changes could indicate that the global climate warming was probably accompanied by the simultaneous global humidization at the end of the Cretaceous.

Reviewers M.A. Akhmet'ev and I.O. Murdmaa

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The Late Riphean Miroedikha Microbiota from Siberia. Part 1: Composition and Facial–Ecological Distribution of Organic-Walled Microfossils

A. F. Veis, P. Yu. Petrov, and N. G. Vorob'eva

Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Abstract—The paper presents new data on taxonomic composition and formation environments of the Late Riphean Miroedikha microbiota from the Turukhansk Uplift, where it consists of organic-walled fossils and is most abundant and diverse as compared to biotas from other localities of the world. Microfossils formerly known from the lower subformation are now recovered from a greater part of the whole Miroedikha Formation. They are encountered at 10 levels of three sections studied at the Miroedikha, Enisei, and Bol'shaya Shorikha rivers. New data considerably extend information about the taxonomic diversity of biota. It includes 52 genera and 78 species of all morphological groups. In addition, nine and seven new morphotypes are established to be classifiable at the species and generic ranks, respectively. Lateral variations in biota composition are analyzed with respect to the transgressive and regressive trends of the Miroedikha basin evolution. The biota is divided for the first time into the shallow-water, transitional, medium-deep "optimal," and deep-water facial-ecological groupings. The established paleoecological localization of some important morphotypes (*Chuaria, Trachyhystrichosphaera, Prolaroforma, Archaeoellipsoides*, and others) was used to correct the formerly accepted standards of their distribution along the paleobasin profile. New data on lateral and vertical distribution of studied microorganisms support our earlier conclusion that their assemblages, displaying the most dynamic evolution and consisting of forms complex in the structure, were localized in the medium-deep part of the shelf, and that evolutionary innovations in the Riphean biota were mostly related to the regressive stages of sedimentary cycles.

Key words: *Riphean, paleobasin, microfossils, transgression, regression, facial-ecological structure of biota.*

INTRODUCTION

One of principal objectives of Precambrian paleontology is to reveal the life-time diversity of most ancient organisms. Considerable progress in solving this problem was made in the period of the 1960s to 1980s, when dozens of rich and diverse microbiotas were discovered and described (for review and references, see Schopf, 1983; Yankauskas *et al.*, 1989; Schopf and Klein, 1992). After the initial euphoria resulting from these discoveries, it became clear that there are many unsolved problems concerning the characteristics of ancient biota and its behavior in space and time. First of all, it was necessary to explain the "asymmetrical" (untypical of the Phanerozoic) composition of biota, whose fossil forms are almost exclusively represented by eu- and prokaryotic producers displaying the low rate of morphological evolution and spontaneous mass appearance of new morphotypes at some discrete stratigraphic levels.

At the same time, the researchers arrived at the conclusion that the aforementioned properties of Precambrian biota are tightly interrelated with and reflect the nontrivial character of interaction with the surrounding medium. To prove this viewpoint, scientists needed a new approach to investigation of ancient microbiotas, such as the reconstruction of ecosystems and their fragments for particular Proterozoic epochs. Unanimously accepting this idea, scientists went separate ways by selecting priorities for the new ecosystem concept. Some of them based their understanding on the uniformitarian hypothesis and attempted to solve the problem using the comparative actualist analysis of Precambrian microbiotas, filling the lack of concrete paleobiological and paleoecological information with conclusions of molecular biology and by *a priori* understanding of hierarchy of biogeochemical cycles in the Earth history (e.g., Krylov and Zavarzin, 1983; Zavarzin, 1984; Knoll, 1992; Schopf and Klein, 1992; Sergeev *et al.*, 1996). Another group of scientists, we are among them, does not discredit the significance of neontological extrapolations, but considers the empirical data on habitat environments of ancient microorganisms as most important for restoring the interaction mode between the Precambrian biota and medium (Vidal and Nystuen, 1990; Zhang and Walter, 1992; Knoll *et al.*, 1991; Hofmann and Jackson, 1994; Veis and Petrov, 1994; Petrov and Veis, 1995; Burzin, 1996; Veis *et al.*, 1998). These data can be obtained only in the course of studies on

distribution of microfossils in deposits accumulated under various conditions, and also by means of correlation between changes in ancient sedimentation settings and concurrent biota. The above consideration shows the key significance of parallel sedimentological and microphytological investigations for modeling the past ecosystems and evolution of Precambrian biosphere.

In this work, we analyze interaction between Proterozoic microorganisms and the surrounding medium using as an example the Miroedikha microbiota (Upper Riphean of the Turukhansk Uplift, Siberia) represented by its organic-walled forms. This microbiota is well known among specialists despite the fact that all published data have been obtained on the basis of few rock samples collected from the lower part the Miroedikha Formation without their detailed positioning in the sampled section (Miroedikha River). Our work in the Turukhansk region provided grounds to enlarge the taxonomic list of this microbiota and to distinguish microfossil assemblages different in composition at ten discrete stratigraphic levels of three sections studied. Using the last model of the Miroedikha shelf evolution (Petrov and Semikhatov, 1997), we considered vertical and lateral variations in composition of these assemblages from the viewpoint of their facial-ecological distribution and stratigraphic succession. As a result, we proved principal trends in distribution of Riphean microorganisms across paleobasins and revealed new peculiarities of microbiota behavior in space and with time, which appear to be of a general significance. The data obtained will be published in two papers. This one (Part 1) summarizes data from the previous work (Petrov and Semikhatov, 1997) about the structure and formation conditions of the Miroedikha Formation, along with new data on taxonomic composition and facial-ecological structure of the Miroedikha microbiota. The next paper (Part 2) will be devoted to the analysis of specific chronological processes of biota evolution and to the first attempt to interpret the succession of Precambrian microorganisms in terms of Phanerozoic paleosuccessions.

MIROEDIKHA FORMATION: STRUCTURE AND ACCUMULATION SETTINGS

The Miroedikha Formation of variegated clay–dolomite deposits occurs in the upper part of the Riphean section of the Turukhansk region. The deposits conformably overlie the Shorikha Formation of dolomites and are subdivided into the lower subformation (37–47 m) mainly composed of dolomite and the upper subformation (120–170 m) representing an alternation of shale and dolomite members (Fig. 1).

The lower subformation includes two members. The first one (12–18 m) consists of gray and pale dolomite beds with thin (2–10 cm) shale interlayers. Dolomites are represented by clayey, horizontally bedded, mechanogenic varieties enclosing interlayers with microbial lamination. Near the base of the member, these deposits incorporate the siliciclastic sandy admixture and show desiccation cracks, erosion marks, and syngenetic brecciation.

The second member (20–28 m) is enriched in clay fraction and contains thick (up to 3 m) interbeds of variegated shales separated by thin transition zones from adjacent beds of fine-grained clayey dolomite with graded bedding. Dolomites of the last type are represented by dolo- and calcarenites in the southern seaward side of the past basin, and by flat-grained intraclastites with coarse cross-wavy and local hummocky bedding, which are typical of the northern proximal part of the basin. The member section is crowned with a persistent bed (3–4 m) of stromatolitic dolomite overlain in the south by a thin sequence (3 m) of clayey dolomites, dolomitic marls, and in situ breccias displaying marks of frequent subaerial expositions. In the north, the last unit is missing, and the top portion of the stromatolitic bed reveals indications of meteoric diagenesis.

The upper subformation is exposed in full measure only in the south of the region. It consists of two members persistent in the lateral direction and of two clay–carbonate members variable in their features. The first member (45–50 m) is composed of uniform carbonate-free shales, dark-gray or sometimes green and red in color, which rest on the basal dolomitic shales (3 m) with grumous structure and desiccation cracks. At the top, the member includes interlayers of clayey–silty siliciclastic and carbonate deposits.

The second, predominantly gray carbonate member (12–32 m) is divided into two bed packets. The lower packet (5 m) consists of alternating shales and fine-grained clastic limestones. The latter show isolated casts of erosion channels and gradually replace shales in the upper part of the section. The upper packet is composed of dolomites. In the eastern exposures, this unit is 7 m thick and consists of horizontally laminated dolosparites with indications of long exsiccation phases. In the west, its thickness grows up to 27 m, and the rocks grade into stromatolitic and clastic carbonate beds with thin shale lenses at the base indicative of the basin deepening.

The third member (57–80 m) includes three packets. The lower well-traceable packet (12–15 m) consists of black pyrite-bearing shales grading upward into dolomitic marls with horizontal bedding. The middle packet is represented in the east by clay–carbonate deposits accumulated in a quiet setting (25–30 m), whereas in the west it consists of clastic and stromatolitic dolomites (up to 50 m) deposited in the course of gradual shoaling. The upper packet (10–12 m) is composed of uniform limestones and dolomites with microbial lamination.

The fourth member (15–18 m) consists of dark gray and black shales, whose lower portion (3–5 m), if compared with the rest of the section, has been deposited in more active hydrodynamic environments and encloses

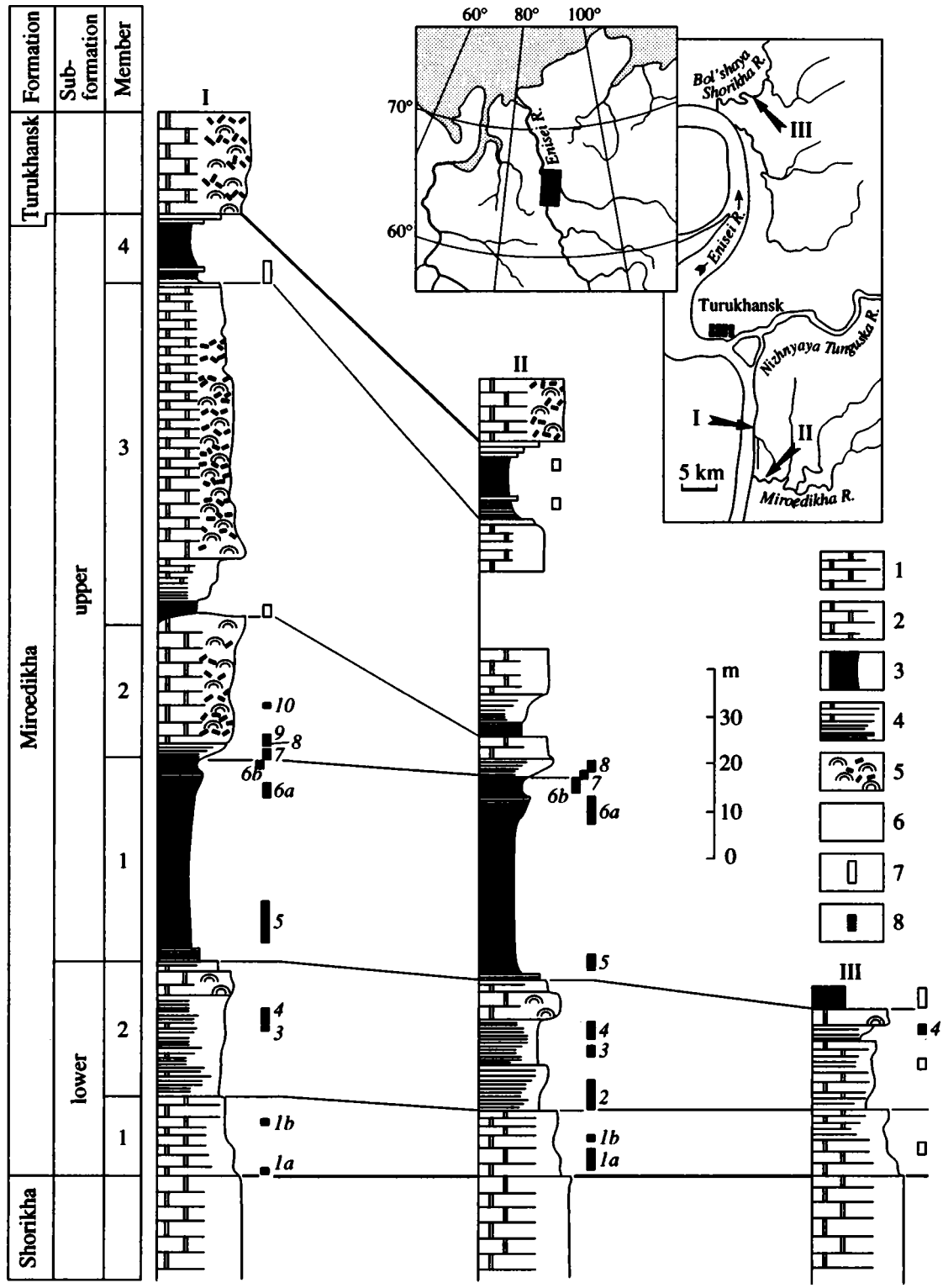


Fig. 1. Geographic position of studied sections of the Miroedikha Formation and their facial composition: (1) limestones; (2) dolomites; (3) shales; (4) intercalated shales and clayey carbonates; (5) clastic limestones and stromatolites; (6) pelitomorphic and fine-grained deposits; (7) sampling interval barren of microfossils; (8) microphytological levels.

dolomitic segregations. This member is conformably overlain by stromatolitic dolomites of the Turukhansk Formation.

All researchers attribute the Miroedikha Formation to the Upper Riphean, though to its different intervals (for review and references, see Petrov and Semikhatov, 1997). Using data of isotopic geochronology and chemostratigraphy, some of them consider the formation as a unit belonging to the lower part of this erathem (within the time span of 1000–850 Ma), while others attribute it to the upper Upper Riphean mainly on the basis of stratigraphic data (microphytological results inclusive).

Deposits of the considered formation were accumulated in the shelf settings. They represent a transgressive rock complex, whose northern section (Bol'shaya Shorikha River) corresponds to the proximal part of the past basin, the distal part of which (Enisei River sections) was situated in the southwest (Petrov and Semikhatov, 1997).

The lower subformation illustrates the gradual establishment of open marine settings and subsequent quick shoaling. Its lower member corresponds to the terminal Shorikha time, when the shallow carbonate platform was flooded to the state of variable settings of the upper subtidal zone with quick exsiccation periods. The upper member, in its main part, corresponds to the mature phase of the transgression with prevailing settings of an open shelf and with indications of the storm-related sedimentation. Its terminal part reflects a quick sea-level drop and appearance of a muddy plain that periodically experienced exsiccation in the distal zone and was completely dried in the proximal shelf area.

The upper subformation corresponds to a more dynamic period of the basin history. The formation time of its lower member marks a quick transgression that created the monotonous environments in an open shelf and was followed by the basin shoaling. The accumulation time of the second member corresponds to the period of considerable differentiation of sedimentation settings in the carbonate shelf of the barrier type. In this period, a shallow lagoon was formed in the inner shelf, and a stromatolitic barrier system appeared in the outer zone. Sediments of the third member suggest the establishment of deep-water anoxic settings of the open shelf and, then, the restoration and next decay of the barrier system with subsequent development of the shallow inner shelf. The fourth member reflects a quick transgression and expansion of settings characteristic of the distal muddy shelf.

The existing model of the Miroedikha basin evolution (Petrov and Semikhatov, 1997) was used to detect the place of the discovered microphytological levels in the main evolutionary trend (Figs. 1 and 2). Levels 1–4 in the first member and lower portion of the second member of the formation are situated within the transgressive sequence of deposits reflecting a transition from tidal and upper subtidal settings to those of the open shelf type. Levels 5–10 are separated from the lower counterparts by a short period of sea-level drop. In general, they are situated within the regressive sequence, which was accumulated after a quick sea-level rise in the region. Level 5 corresponds to deposits accumulated in a stable deep-water setting, and levels 6–10 depict the trend of shoaling, in other words, a transition from an open muddy shelf to the lower subtidal zone of the clay-carbonate sedimentation and then to the upper subtidal zone with the sandy carbonate shoals. Environmental characteristics of each microphytological level are presented below.

ORGANIC-WALLED MICROFOSSILS FROM THE MIROEDIKHA FORMATION

Material and investigation methods. Microfossils were discovered from the Miroedikha Formation by Timofeev (1969) and immediately attracted attention due to the abundance and perfect preservation degree of sphaeromorphic sheaths typical of the Precambrian and associated with organic-walled colonial coccoidal and filamentous remains poorly known at that time, and with fossils of complex morphology similar to algae and mycobiota (German, 1974; Timofeev *et al.*, 1976). Many forms described later from this locality (for references, see Veis, 1988; German, 1990) were of a great paleobiological and stratigraphic significance and played the key role in destroying a barrier that separated scientists specialized in organic-walled and mineralized microfossils (Yankauskas *et al.*, 1989). At the same time, all attempts to find microfossils at other levels, or in other sections, ended with the discovery of impoverished assemblages of unimportant forms (Mikhailova, 1987; German, 1990). This situation resulted in the illusion that the Miroedikha microbiota is unique and appeared, as it was assumed (Timofeev *et al.*, 1976; German, 1990), due to the unusual habitat environments and/or fossilization. Inferences concerning these environments were limited by speculations about origin and burial of the biota in a quiet and shallow marine basin with normal or slightly lowered salinity (Ivanovskaya *et al.*, 1979) that was, of course, a vague explanation for the biota uniqueness.

Our investigation of 1978–1995 considerably extends the data base on composition and geographic and stratigraphic ranges of microremains from the Miroedikha locality. New data allow us to consider for the first time the paleoecological and chronological aspects of the discovered assemblages different in taxonomic composition. We studied 93 rock samples collected from three sections (Fig. 1) located at the Miroedikha River (48 samples), at the Enisei River downstream from the Kamennaya River mouth (40 samples), and near the Bol'shaya Shorikha River mouth (5 samples). In the last locality, only the lower part of the formation is exposed. In some cases, we also collected samples with little promise in the phytological aspect but useful for a complete understanding of facial distri-

bution of microfossils. We discovered microfossils in a sufficient or perfect preservation state from 53 samples (31, 20, and 2 samples, respectively, from each of the indicated localities, or 56% of the total collection).

The procedure of the laboratory treatment of microphytological samples and the accepted classification of microfossils are described in our previous publications (Veis, 1988, 1993; Veis and Vorob'eva, 1992, 1993; Veis and Petrov, 1994). In distinction to other classifications, we take into account only the real (observable) morphological features and reduce to a minimum the interpretation of their possible biological significance (for instance, see Knoll *et al.*, 1991; Butterfield *et al.*, 1994; Sergeev *et al.*, 1995) excluding from consideration all morphological details secondary in origin. In such a manner, we distinguish the broad morphological groups of the supergeneric rank like acritarchs, colonial coccoidal, filamentous (sheaths and trichomes) and complicated forms, and also the authorized taxa of generic and species levels (Yankauskas *et al.*, 1989, with minor modifications).

Prior to our investigation, 38 genera and 48 species had been detected in the Miroedikha biota. According to our data, it includes 52 genera and 78 species described in various publications and associated with a series of new indicative taxa. Acritarchs (14 genera, 25 species, 40% of all microfossils) and filamentous sheaths (14 genera, 16 species, 25% of all forms) are leaders in the taxonomic diversity. They are followed by trichomes (9 genera, 18 species, 12% in the general composition) and colonial coccoidal microfossils (8 genera, 12 species, 15% of all specimens). Remains of a complex morphology are less abundant than others (7 genera, 7 species, 8% of the assemblage). Amid the morphotypes not yet described, there are acritarchs, filamentous and complex forms (at least 7 new genera and 9 new species). As for the taxa previously reported from the Miroedikha locality (German, 1990), our material is lacking the acanthomorphic (?) acritarchs *Miroedichia* Herm. known in single specimens, and also the problematic branching thalluses *Palaeocalothrix* Herm., the latter discredited as *in situ* form (Hofmann and Jackson, 1994).

Distribution of microfossils through the section and region. Microfossils are detected within ten microphytological levels of the Miroedikha Formation, i.e., in the thin (few meters) intervals of the section saturated with microremains and separated by larger intervals barren of the latter. In some cases, when the lower and upper portions of a level turned out to be different in composition of microfossils, we distinguished them as the individual sublevels. The interregional correlation of the levels is based on lithological and stratigraphic criteria, and sometimes on specific combinations of morphotypes characteristic of the particular levels. These combinations are a sort of "visit card" for such levels.

Discrimination of microphytological levels is the necessary element of our approach to investigation of Precambrian microbiotas, because it reveals positions of discrete microfossil assemblages in a given section and is useful for considering their localization in the sedimentological profile of the past basin. Another important element of this approach is the concept of a limited posthumous transportation of Riphean microorganisms that distorts the original patterns of their facial distribution. We do not mean the strictly autochthonous status of all biota members, but suggest that they have not been displaced from their original and sufficiently wide facial-ecological zones for a distance of statistical significance. This concept was substantiated in our previous publication (Veis and Petrov, 1994).

Most rich and diverse assemblages of microfossils are detected in the Miroedikha (levels 1–8) and Enisei (levels 1 and 3–10) sections. In the Bol'shaya Shorikha section, the representative assemblage was recovered from Level 4 only (Fig. 1). On the basis of morphology and according to lateral and vertical distribution ranges, all recognized microfossils can be divided into two categories of unequal significance. The first category includes the most widespread and simple morphotypes characteristic of the entire Riphean sequence. These are acritarchs *Leiosphaeridia* (10–300 μm),[1] colonial coccoidal *Myxococcoides* (8–16 μm), sheaths *Eomycetopsis* (2–4 μm) and *Leiotrichoides* (5–16 μm), and trichomes *Oscillatoriopsis* (4–8 μm). These taxa are dominants at all ten levels, explicate the microphytological characteristics of some of them, and are indifferent (except the quantitative variations) with respect to paleoenvironments. In the further description, their presence will not be mentioned unless they are the only representatives of the particular level.

Remarkable morphotypes of much more limited stratigraphic ranges are referred to the second category. These are the Late Cambrian *Chuaria*, *Valeria*, *Prolatoforma*, *Trachyhystrichosphaera*, *Ostiana*, *Polytrichoides*, *Taenitrichoides*, *Arctacellularia*, *Trachytrichoides*, *Elatera*, and others known in many regions, and also the interregional taxa *Cerebrosphaera*, *Pseudodendron*, *Cephalonyx*, *Polysphaeroides filiformis* Herm., *Palaeolyngbya sphaerocephala* Herm. et Pyl., *Eosolena*, and *Arthrosiphon*,[2] and endemic *Miroedichia* along with all new genera and species. Morphotypes of the second category are dispersed in deposits more irregularly than those of the first category.

[1] All size ranges are presented for most characteristic morphotypes of the Miroedikha biota. They indicate the sheath diameters for acritarchs, sheath and trichome width for the filamentous forms, dimensions of shorter and longer axes for the ellipsoid bodies, maximum sizes of main morphological elements for the fossils of a complex structure, and diameters of individual cells of colonial coccoidal forms; dimensions of colonies and their aggregates are shown after a semicolon.

[2] Tubular forms displaying characteristic bulges and belonging to the genus *Arthrosiphon* A. Weiss have been erroneously attributed by Butterfield *et al.* (1994) to the genus *Cephalonyx* A. Weiss incorporating different morphotypes.

They show distinct relations to concrete paleosettings that existed during the transgressive and regressive trends of the basin evolution and at the initially stable deep-water stage of the second trend. Most characteristic morphotypes associated with both trends are illustrated in Fig. 2. Lateral variations in their sizes are presented in Fig. 3, and images are shown in Plates I–III.

The transgressive stage of the basin evolution corresponds to the lower subformation comprising four lower microphytological levels

First level is subdivided into the lower and upper sublevels with rather different microfossil assemblages. The former (1.5–4.5 m above the formation base at the Miroedikha River) yielded only rare microfossils of the first category. In the upper one (8.5–9 m above the formation base at the Miroedikha River and 13 m above it at the Enisei River), the same forms are more abundant and coexist with aggregated sheaths of *Leiotrichoides* and with dispersed sheaths of multiserial *Polytrichoides* (4 μm) and glomerular *Glomevertella* (6 μm). Microfossils are recovered from interbeds of dark green shales variable in thickness (5–20 cm) and intercalated with more thick beds of clayey, horizontally laminated dolomites. At the Miroedikha River, these rocks display marks of a brief subaerial exposition. The deposits accumulated in uniform, predominantly quiet (protected) settings of the upper subtidal or tidal (Miroedikha River section in the east of the region) zones. It is remarkable that microfossils are more diverse in the deeper shelf deposits (the upper sublevel at the Enisei River) than in sediments of shallower settings (Miroedikha River).

Second level (0–6.5 m above the base of the second member of the lower subformation, the Miroedikha River) includes rare and unimpressive microfossils of the first category, which are recovered from the rather thick (up to 0.5 m) beds of gray shales intercalated with those of clayey dolomites and doloarenites. These sediments accumulated in the laterally persistent shallow settings of the upper subtidal zone that gradually deepened with time. Doloarenites, which should be accumulated close to the basis of the constant wave activity, suggest the seaward progradation of sedimentation setting as compared to those of the first level. It is also remarkable that sediments of the second level exposed at the Enisei River, where they are barren of microfossils, show ephemeral signs of accumulation under the stagnant conditions.

Third level (12–14 m above the base of the second member of the lower subformation at the Miroedikha River and 14–16 m above it at the Enisei River) encloses the abundant cenobial aggregates of small *Leiosphaeridia* (20–30; 60–100 μm) and forms with internal bodies attributable to *Pterospermopsimorpha* (30–50 μm), giant *Chuaria* (1200 μm), and sheaths of *Polytrichoides* (3–6 μm). In addition, we encountered here the following morphotypes: large forms of *Leiosphaeridia* (up to 300 μm) and *Chuaria* (up to 700 μm), including *Ch. globosa* Serg. et Ogurtz. (up to 350 μm); oval *Cucumiforma* with the longitudinal structural patterns (120 × 240 μm); spheroid colonies of *Sphaerophycus* (7; 30–40 μm); mono- and biserial *Sphaerocongregus* (16–18 μm in aggregates as long as 100 μm); thalluses of *Ostiana* (10–12; hundreds μm); empty subspheroidal colonies of *Coneosphaera* (42–48; 250–300 μm); glomerular sheaths of *Glomovertella* (4–6 μm); ellipsoidal *Archaeoellipsoides* (10 × 40 μm); aggregates of thin filamentous *Eomicrocoleus* (1 μm); trichomes of *Filiconstrictosus* (5 and 7 μm); *Oscillatoriopsis* (up to 20 μm); *Cephalophytarion* (4–5 and 12 μm); and *Palaeolyngbya* (14–40 μm), including *P. sphaerocephala* (40 μm) with a fine transverse ornamentation. Other characteristic forms are represented by heterocyte trichomes (2 × 4μm) with considerably enlarged cells (6 × 10–12 μm), by trichomelike *Arctacellularia* (20 μm), and by complex morphotypes such as branching and nonbranching thalluses with a vague longitudinal structure (10 and 50 μm) resembling *Pseudodendron*.

The listed morphotypes occur in thick (up to 2 m) beds of gray to dark green shales intercalated with beds of clayey dolomites. These deposits are lacking indicators of their accumulation under influence of wave activity and primarily reflect episodes of sedimentation events. In other words, deposits of the third level accumulated in a deeper setting of the lower subtidal zone as compared to sediments of the second level.

Fourth level (16–19 m above the base of the second member of the lower subformation at the Miroedikha River, 16–20 m above this datum at the Enisei River, and 18–19 m above it at the Bol'shaya Shorikha River) yielded a very diverse assemblage of abundant microfossils. Most remarkable among them are the following morphotypes: acanthomorphic acritarchs *Trachyhystrichosphaera* (140–200, sometimes up to 800 μm); *Prolatoforma* (200 × 250 μm); *Chuaria* of medium size (up to 600 μm), including *Ch. globosa* (450 μm); marginate *Simia* (190–450 μm); sheaths of *Polyspaeroides filiformis* Herm. with the transverse series of inner bodies (3–5; 40 μm); multiserial *Polytrichoides* (4–6 μm); trichomes *Cephalophytarion* (2–12 μm); *Filiconstrictosus* (6–9 μm); and "multichamber" *Eosolena* of a complicated structure (60–90; 100 × 150 μm). These morphotypes coexist with abundant monstrous acritarchs and their aggregates (3 to 8 envelopes up to hundreds μm in diameter each), and also with lophostriated *Valeria* (300–330 μm), sheaths of *Cerebrosphaera* with "brain" wrinkles at the surface (200–210 μm), ellipsoidal *Konderia* (160 × 280 μm), prismatic *Octaedrixium* (20–22 μm), and with very large *Pterospermopsimorpha* and *Nucellosphaeridium* (more than 500 μm). Filamentous forms (sheaths, trichomes, and trichomelike morphotypes) are also important components of this assemblage. The smooth-walled sheaths (2–100 μm) lacking ornamentation are represented by *Eomycetopsis* (2–4 μm) and *Leiotrichoides* (6–20 μm) composing the dense mats, and also by *Asperatofilum* (40–50 μm),

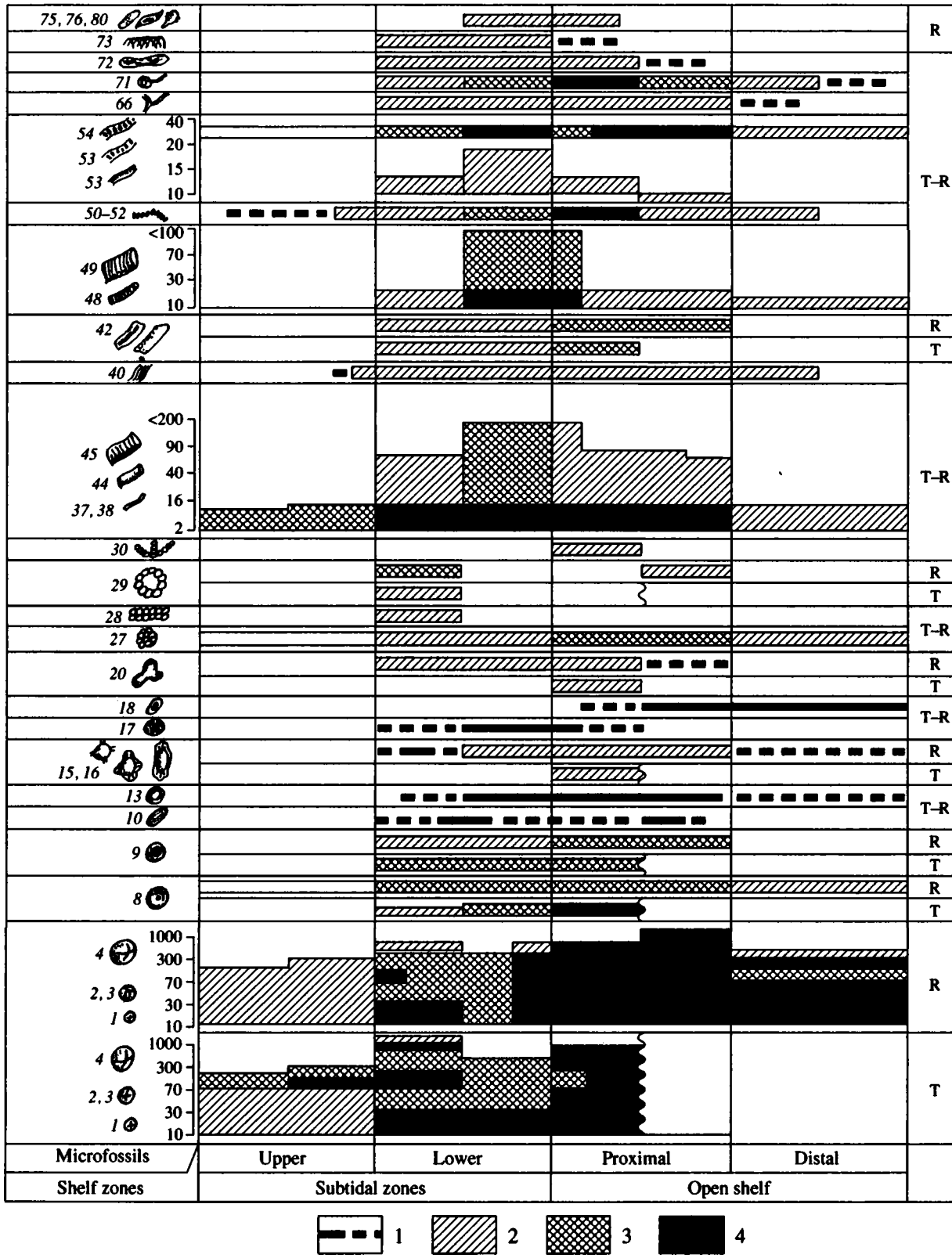


Fig. 2. Dimensional characteristics (1st, 2nd, 20th, 24th, and 26th diagrams from below, scales in μm) and occurrence frequency (other diagrams) of most indicative morphotypes and their relations with principal zones of the Miroedikha shelf: (1) occasional finds; (2) single specimens; (3) usual components; (4) abundant microfossils (numbered drawings are taxa listed in explanations to Fig. 3).

Taenitrichoides (70–100 μm), and *Entospha-eroides* (18 and 36–40 μm), the latter displaying the scarcely dispersed internal bodies. Among sheaths with the transverse structure, predominant forms are *Siphonophycus* (10–26 μm) with the narrow dark costae and light zones displaying the former cellular structure of decayed trichomes, wide *Plicatidium* (90–130 μm) with the thin frequent striation, and wrinkled-costate *Rectia* (38–40 μm). They associate with glomerular *Glomovertella* (4–6 μm), small spiral *Obruchevella* (2 μm, spiral diameter 6–7 μm), and ellipsoidal envelopes of *Archaeoellipsoides* (8–30 × 30–200 μm) conventionally attributed to the sheaths. Many of these envelopes enclose the dark trichomelike bodies thus being very similar to akinetes of hormogonian cyanobacteria (Sergeev *et al.*, 1995). Some of them are lacking the internal inclusions and display a vague tangential striation. Among the transitory and aforementioned characteristic filamentous forms, we identified the following taxa: trichomes of *Calyptothrix* (4–12 μm);[3] *Botuobia* (16–20 μm); *Palaeolyngbya* (8–20 μm), *P. sphaerocephala* (40 μm); *Cyanonema* (3 μm); heterocyte morphotypes, whose cells are unusually large (from 4–8 × 10 to 12–14 × 20 μm); trichomelike *Arthrosiphon* (4–6 μm); *Trachytrichoides* (20–25 μm); and large articulate *Cephalonyx* (48 μm).

Microfossils of a complex structure are represented at the fourth level by thalluses of *Pseudodendron* (40–100 μm), by *Germinosphaera* envelopes with the solitary processes (60 and 90 μm, processes up to 10–30 μm) by *Caudosphaera* (up to 400–500 μm, processes up to 100 μm), and by irregularly fusiform *Fabiformis* (50–60 × 300–600 μm) cancellate in structure and growing wider at one or both ends. In addition, the assemblage includes three morphotypes not yet described: (1) thin infundibular membranes (70–90 μm), which may represent either the whole organisms, or the fragments of more complex structures; (2) elongated and curved aggregates (140–200 μm) of spheroid bodies (46–50 μm) with a budlike termination (90 μm); and (3) solitary C-shaped aggregates of thin sheaths (1–2; 290 μm). Thalluses of *Ostiana* (12–14 μm; hundreds μm long), *Polysphaeroides* (2–3; 13–14 and 40 μm), and rare pseudofilamentous *Palaeopleurocapsa* (10–12; 40–60 μm) are the most remarkable morphotypes among abundant and diverse colonial coccoidal microfossils.

The listed microfossils are recovered from dark gray to greenish shale beds (up to 3 m in thickness) intercalated with clayey dolomite and marl beds. These deposits typical of the proximal open shelf settings are widespread almost everywhere in the region. The graded bedding is their characteristic feature in the southern sections, where it is seen at the base of thick (up to 0.8 m) carbonate beds and marks a transition from the fine-grained shale breccias with the clay–carbonate matrix to the fine-grained, and then to pelitomorphic dolomites. This kind of bedding suggests a brief erosion episode followed by precipitation of sedimentary material from the turbidity currents, and, accordingly, we consider these deposits in the southern areas as the distal tempestites. In the north, the lower subtidal zone of the inner shelf (Bol'shaya Shorikha River) accumulated the carbonate–clay sediments under conditions of the persistent and relatively active hydrodynamics. The lateral differentiation of the open shelf settings was clearly responsible for the observed compositional changes in microfossil assemblages of the fourth level: all acanthomorphic acritarchs disappeared from the inner shelf; the diversity of trichomes, and of complex morphotypes especially, drastically decreased, but very wide smooth-walled tubercular microfossils became rather abundant. Moreover, we should attract attention to the abundance of akinetelike forms discovered from the proximal shelf deposits, in contrast to the recent conclusion (Knoll and Sergeev, 1995) that they predominantly occur in the tidal zones of the Early Riphean carbonate platforms.

The regressive stage in the basin evolution is recorded in two lower members of the upper subformation, where we detected all other levels enriched in microfossils. The greater part of the first of them characterizes the stable deep-water environments of sedimentation at the beginning of the stage (Petrov and Semikhatov, 1997) and reveals, as we will see below, an obvious microphytological peculiarity of the fifth level.

Fifth level (3–6 m above the base of the upper subformation at the Miroedikha River and 5–14 m above it at the Enisei River) yielded an assemblage of poorly diverse taxa. Its characteristic morphotypes are only colonies of coccoidal microfossils: shapeless *Myxococcoides* (8–12; 30–60 μm) and spheroid *Sphaerophycus* (8–12 and 16–18; 30–50 μm). In addition, the assemblage includes the relatively abundant *Leiosphaeridia* (up to 300 μm), less frequent *Chuaria* (300–500 μm), *Ch. globosa* (350 μm), fragmented *Trachyhystrichosphaera* (200), cancellate *Spumosina* (50 μm), and ellipsoidal envelopes (150 × 200 μm) resembling *Navifusa*. Filamentous forms are represented by some transitory taxa and by rare *Polytrichoides* (4–6 μm) and *Palaeolyngbya* (15–40 μm), *P. sphaerocephala* (40 μm) inclusive. Representatives of the complex morphotypes are *Caudosphaera* (100–120 μm, processes more than 100 μm long) and accumulations of narrow *Pseudodendron* thalluses (10–12 μm). Many samples bearing microfossils are enriched in the decomposed organic matter suggesting the former existence of abundant benthic coenoses of destructive microorganisms.

The listed microfossils are concentrated in the uniform clayey sediments deposited in the most quiet and deep settings of the distal open shelf located below the storm wave basis and outside the ebb-current zone. Microstructures and composition of these sediments suggest the steady-state character of sedimentation.

[3] Trichomes of genera *Cephalophytarion* Schopf and *Calyptothrix* Schopf have been attributed by German (1990) to genus *Primorivularia* Edhorn.

Fig. 3.

Sixth level consists of two sublevels with different microfossil assemblages. Its lower sublevel (35–42 m above the base of the upper subformation at the Miroedikha River and 37–40 m above it at the Enisei River) is noticeable due to presence of giant *Chuaria* (up to 1200 μm), *Polytrichoides* sheaths (2–4 μm), filamentous *Palaeolyngbya sphaerocephala* (40 μm), *Pseudodendron* thalluses (20–25, 40, and 50 μm), *Eosolena* (100 μm), and several complex forms unknown previously. We also identified here acritarchs *Leiosphaeridia* (up to 300 μm), lesser *Chuaria* (up to 800 μm), *Trachyhystrichosphaera* whose very large forms (up to 800 μm in diameter) have the well-developed membranous margin, *Prolatoforma* (180–260 μm), striated

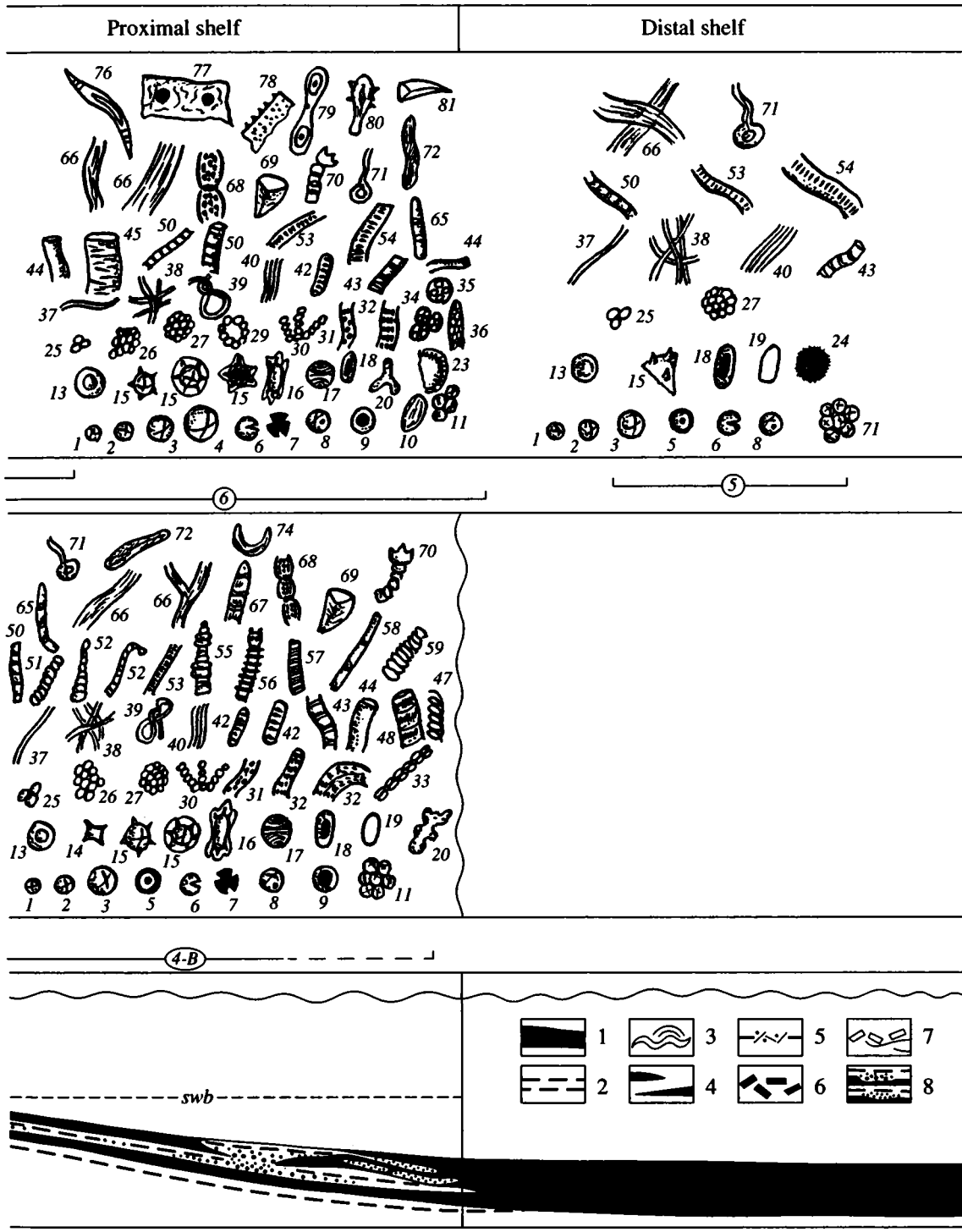


Fig. 3. (Contd.)

Cucumiforma (60 × 160 μm), and ellipsoidal *Konderia* (140 × 200 μm). Colonial coccoidal forms are represented by thalluses of *Ostiana* (10–12; hundreds μm), *Polysphaeroides* (1–2; 24-40 μm) enclosed into the filamentous sheaths, spheroid *Sphaerophycus* (4; 20–30 μm), subspheroidal *Eogloeocapsa* (12–14; 30 μm), and empty colonies of *Coneosphaera* exemplified by two generations different in size (10; 100–110 μm, and 30; 480–500 μm). Filamentous forms of this sublevel include: *Eomycetopsis* (2–4 μm) and *Leiotrichoides* (5–16, sometimes up to 30 μm), both occurring either as abundant disjoined specimens or as constituents of

Fig. 3. Paleontological characteristics of distinguished microphytological levels and their position relative to the Miroedikha shelf: (1) shales; (2) fine-grained carbonates; (3) microbioliths; (4) thin interlayers and lenses of shale in carbonate deposits; (5) broken carbonate interlayers and fine-grained carbonates; (6) shale breccia; (7) coarse-grained carbonates; (8) coarse- to fine-grained graded bedding. Bars with encircled numbers denote facial environments of studied microfossil assemblage from the Bol'shaya Shorikha (*A*) and Miroedikha–Enisei (*B*) sections; letter symbols "fwb" and "swb" denote basis of wave activity during the ordinary and storm weather, respectively.

Numbered drawings denote the microfossil taxa: *1*, small- to medium-sized *Leiosphaeridia* (*Leiominuscula* inclusive); *2*, large *Leiosphaeridia*; *3*, small- to medium-sized *Chuaria*; *4*, large *Chuaria*; *5*, *Ch. globosa* Ogurtz. et Serg; *6*, *Leiosphaeridia bicrura* Jank.; *7*, *L. ternata* (Tim.); *8*, *Nucellosphaeridium* spp.; *9*, *Pterospermopsimorpha* spp.; *10*, *Cucumiforma vanavaria* Mikh. emend. Mikh.; *11*, cenobial aggregates of small *Leiosphaeridia*; *12*, *Cerebrosphaera* sp.; *13*, *Simia nerjenica* A. Weiss and *Simia*-like envelopes; *14*, *Octaedrixium* sp.; *15*, *Trachyhystrichosphaera* spp. (marginate and immarginate envelopes); *16*, *Prolatoforma aculeata* Mikh.; *17*, *Valeria lophostriata* Jank.; *18*, *Konderia elliptica* A. Weiss; *19*, *Navifusa*-like envelopes; *20*, monstrous acritarchs; *21*, *Satka* sp.; *22*, envelopes ex gr. *Valeria* with the hypertrophied lophostriation; *23*, *Cymatiosphaeroides* sp.; *24*, *Spumosina* sp.; *25*, *Myxococcoides* spp.; *26*, *Ostiana microcystis* Herm.; *27*, *Sphaerophycus* sp.; *28*, *Sphaerocongregus variabilis* Moor.; *29*, *Coneosphaera arctica* Hofm.; *30*, *Palaeopleurocapsa* sp.; *31*, *Entosphaeroides* sp.; *32*, *Polysphaeroides filiformis* Herm.; *33*, *P. contextus* Herm.; *34*, *Coniunctiophycus* spp.; *35*, *Eogloeocapsa* sp.; *36*, *Chlorogloeopsis* sp.; *37*, *Archaeotrichion contortum* Schopf, *Eomycetopsis robusta* Schopf emend. Knoll, Golubic, *Leiotrichoides typicus* Herm. emend. Herm. and/or *L. gracilis* Pjat.; *38*, mats consisting of sheaths of the same species; *39*, *Glomovertella eniseica* (Herm.); *40*, *Polytrichoides lineatus* Herm. emend. Herm.; *41*, *Eomicrocoleus* crassus Horod. et Donald.; *42*, *Archaeoellipsoides* spp.; *43*, *Siphonophycus* spp.; *44*, *Asperatofilum experatus* (Herm.) and/or *Taenitrichoides jaryschevicus* Ass.; *45*, *Plicatidium latum* Jank.; *46*, *Rectia costata* (Jank.); *47*, *Obruchevella* sp.; *48*, narrow or medium in size cuticular (?) morphotypes with the transverse structural patterns; *49*, wide cuticular (?) morphotypes with the transverse structural patterns; *50*, *Oscillatoriopsis* spp.; *51*, *Filiconstrictosus* spp.; *52*, *Cephalophytarion* spp.; *53*, narrow and medium in size *Palaeolyngbya*; *54*, *P. sphaerocephala* Herm. et Pyl.; *55*, heterocyte trichomes; *56*, *Arthrosiphon* spp.; *57*, *Botuobia* spp.; *58*, *Cyanonema* spp.; *59*, *Calyptothrix perfecta* A. Weiss; *60*, problematic remains of bangiophytes; *61*, *Circumvaginalis elongatus* Serg.; *62*, *Rugosoopsis tenuis* Tim. et Herm.; *63*, large trichomelike morphotypes; *64*, *Arctacellularia* spp.; *65*, *Trachytrichoides ovalis* Herm.; *66*, size-variable thalluses with vague longitudinal structure ex gr. *Pseudodendron* and their interlaced aggregates; *67*, *Cephalonyx sibiricus* A. Weiss; *68*, *Eosolena anisocyta* Herm.; *69*, rounded-triangular membranous morphotypes; *70*, sublinear aggregates of cells with budlike terminal structures; *71*, *Germinosphaera* spp. and/or *Caudosphaera expansa* Herm. et Tim.; *72*, *Fabiformis porosus* Pjat.; *73*, *Elatera binata* Herm.; *74*, C-shaped remains with fibrous longitudinal texture; *75*, large elongated envelope with problematic multichamber structure; *76*, a giant form ex gr. *Pellicularia*; *77*, problematic thalluses of Metaphyta; *78*, laminated crenulate fragments of organisms whose morphology is unknown; *79*, *Pseudotawuia* sp.; *80*, clavate morphotypes with conic processes in their wider part; *81*, narrow acute-triangular morphotypes resembling *Ceratophyton*.

dense mats; smooth-walled *Asperatophillum* (40–44 μm); *Entosphaeroides* (1–2; 17–20 μm) with dispersed internal bodies; glomerular *Glomovertella* (4–6 μm); and abundant akinetelike *Archaeoellipsoides* (6 × 70 to 16 × 100 μm). Trichomes are outsiders of this level being represented only by transitory forms such as *Palaeolyngbya* (12–20 μm) and *P. sphaerocephala* (40 μm).

At the sublevel under consideration, microfossils of a complex structure include seven remarkable morphotypes. Three of them: *Fabiformis* (30 × 300 to 50 × 530 μm), sublinear aggregates of spheroid bodies with budlike terminations (60 × 120 μm), and *Caudosphaera* (30–70 μm, processes up to 200 μm long), also occur at the previous level, but others have nor been encountered below. These are: (1) very large *Pellicularia* (up to

Plate I. Microfossils from the Upper Riphean Miroedikha Formation of the Turukhansk Uplift:

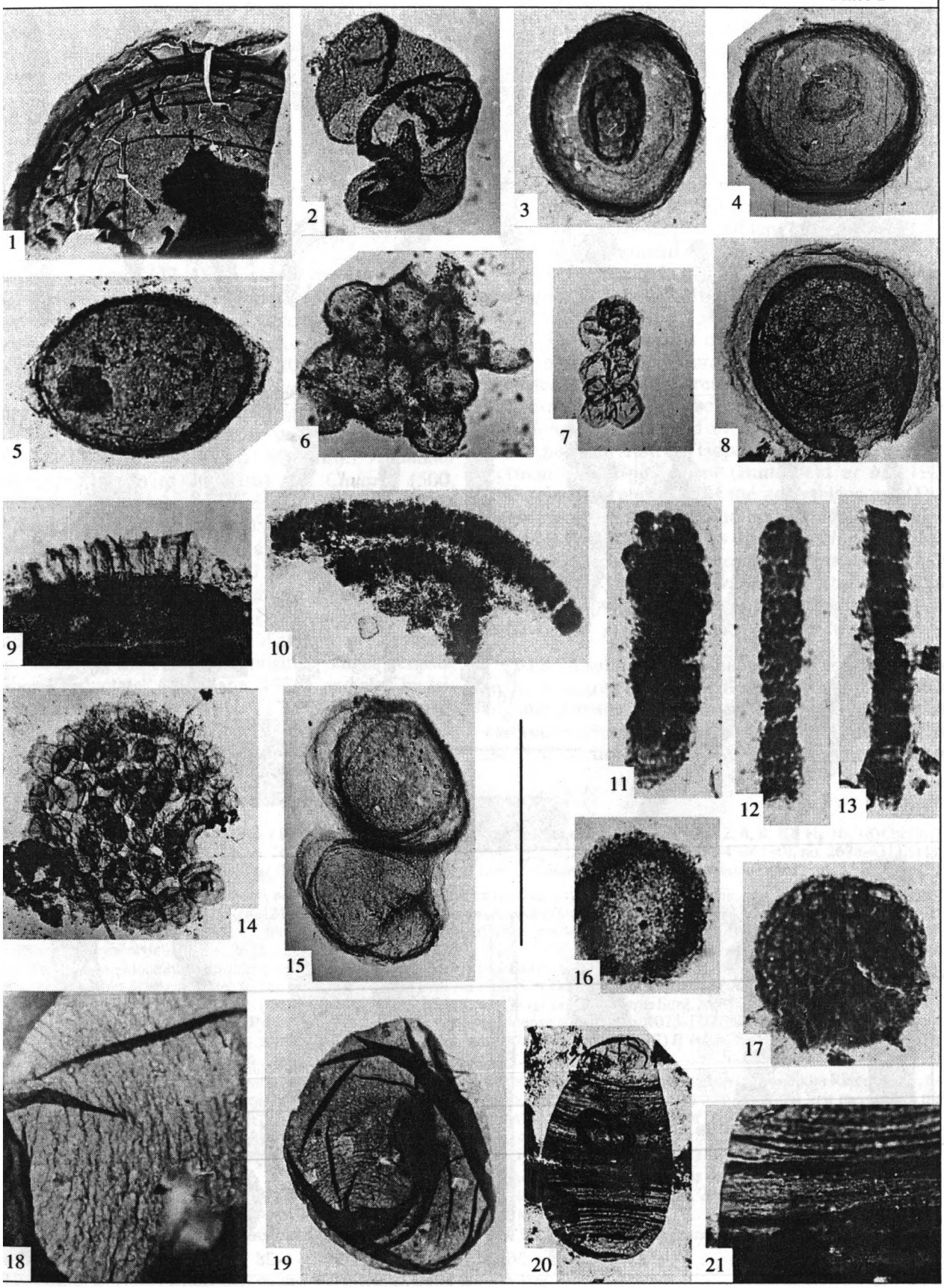
(1) *Trachyhystrichosphaera aimika* Herm., no. 907-720; (2, 15) envelopes of irregular shape (monstrous acritarchs) ex gr. *Leiosphaeridia*, no. 2578-4210 (2) and ex gr. *Simia*, no. 2578-6370 (15); (3, 4, 8) *Simia nerjenica* A. Weiss, marginate envelopes with peripheral plicae and inner bodies different in shape and size, no. 2578-6497 (3), no. 1407-300 (4), no. 4085-700 (8); (5) *Prolatoforma aculeata* Mikh., ellipsoidal envelope with a rim and processes at its poles, no. 4085–500; (6) *Myxococcoides grandis* Horod. et Donald., a colony of cells with inner bodies representing remains of the collapsed cytoplasm, no. 2578-4200; (7) *Sphaerocongregus variabilis* Moor., biserial aggregate of coccoidal cells, no. 2678-4399; (9) *Cymatiosphaeroides* sp., envelope with abundant and thin curved acicular processes immersed in the multilayered(?) filmy rim, no. 4082-400; (10–13) *Polysphaeroides filiformis* Herm., a fragment of unusual aggregate with subconcentric structure, no. 2678-4250 (10), no. 2678-4208 (11), no. 2578-6126 (12), no. 2678-4202 (13); (14) *Coneosphaera arctica* Hofm., empty subspheroid colony with coccoidal cells along the periphery, no. 1427-100; (16) *Simia* sp., no. 4085-990; (17) *Sphaerophycus* sp., dense aggregate of coccoidal cells deformed in response to compression, no. 2678-4225; (18, 19) *Valeria lophostriata* Jank., sphaeromorphic envelope with the fine striation distinct under magnification (18), no. 2678-4460 (19); (20, 21) pyriform envelope with the hypertrophied lophostriation ex gr. *Valeria*, no. 907-1712 (20), central part of the envelope under magnification (21).

Scale bar size: 130 μm for image 1; 80 μm for images 2, 3, 9, 11; 300 μm for images 4 and 19; 250 μm for images 5, 16, 20; 100 μm for images 6, 12, 18, 21; 160 μm for image 7; 120 μm for images 8, 13, 17; 280 μm for images 10 and 14; 400 μm for image 15.

The occurrence levels of microfossils: (14) third level, the second member of the lower subformation (Miroedikha River); (7, 10–13, 17–19) fourth level (ditto); (1, 4) fifth level, the first member of the upper subformation at the Enisei (1) and Miroedikha(4) rivers; (3, 15) lower sublevel of the sixth level, the first member of the upper subformation at the Miroedikha River; (9) upper sublevel of the sixth level (ditto); (5) seventh level, the second member of the upper subformation at the Miroedikha River; (2, 6, 8, 16, 20, 21) eighth level, the second member of the upper subformation at the Miroedikha (2, 6, 8, 16) and Enisei (20, 21) rivers.

All specimens whose images are presented are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

Plate I



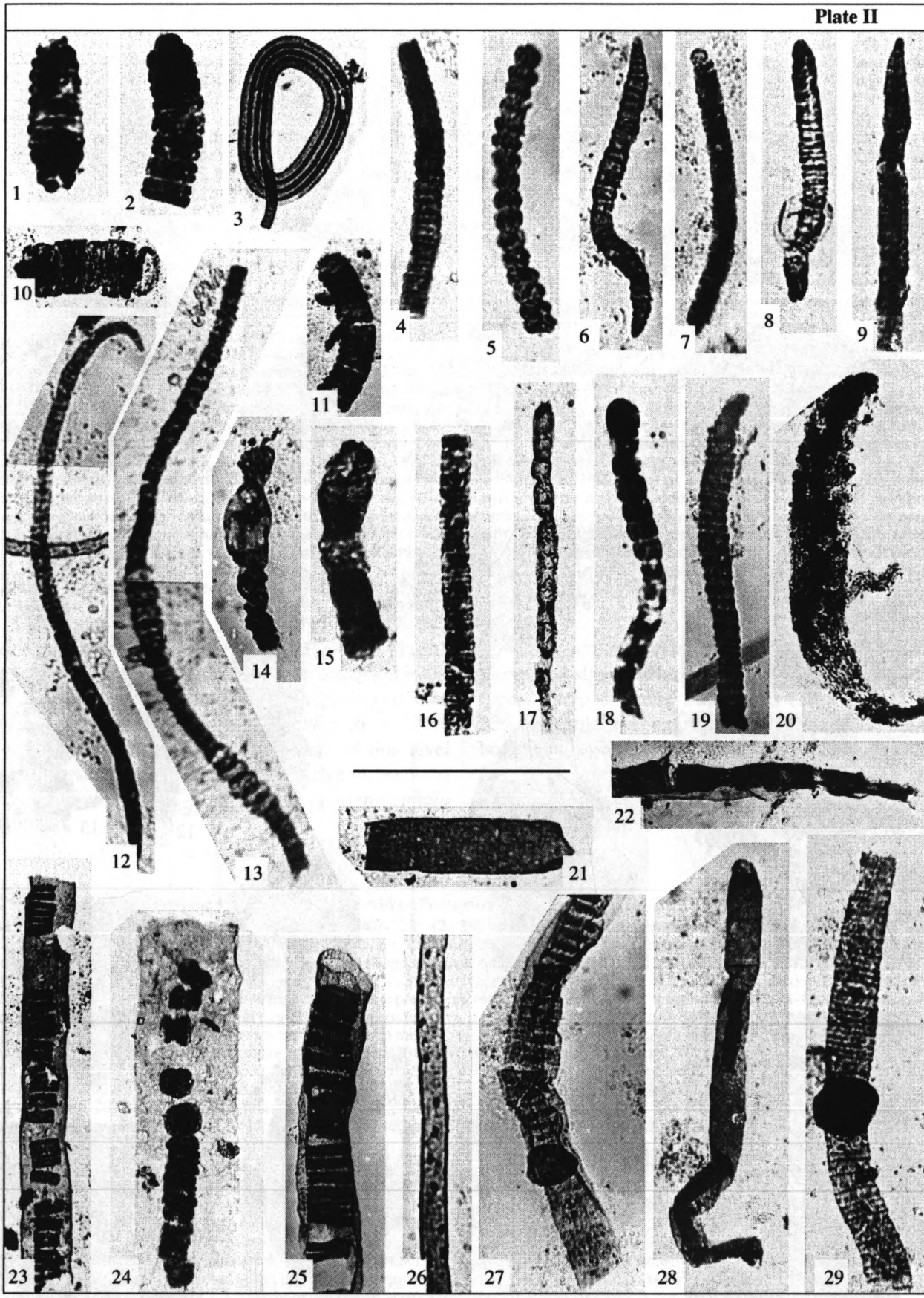


Plate II
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1600 μm) displaying the distinct multicellular(?) structure of acute terminations and the well-outlined longitudinal rima previously unrecorded as a feature of the genus; (2) thin acute-conical membranes (up to 200 μm); (3) very wide thalluses (wider than 500 μm) resembling the Vendian Metaphyta, enclosing rounded "sporangia" (to 100–150 μm) in their central part, and decorated with abundant and small rodlike elements (2–3 × 8–10 μm), which are similar to the fossil actinomyces *Primoflagella*; and (4) elongated (30–40 × 100–150 μm) lamellar fragments of some larger organisms, which are ornamented with short spinules.

Upper sublevel (42–45 m above the base of the upper subformation at the Miroedikha River and 44–46 m above it at the Enisei River) yielded the indicative assemblage of monstrous acritarchs (up to 350 μm), aggregated *Coniunctiophycus* (10–20; 100 μm, and 40–60; 160 μm), very thin (1 μm) *Archaeotrichion* sheaths composing mats, filamentous *Taenitrichoides* (80–90 μm) and *Plicatidium* (120–260 μm), and *Pseudodendron* thalluses (25, 50, 70, and 90 μm). Abundant *Leiosphaeridia* (up to 300 μm) and *Chuaria* (500, sometimes 700 μm), lophostriated *Valeria* (500–600 μm), marginate *Simia* (100 μm), acanthomorphic *Cymatiosphaeroides* (fragments up to 60 μm), and ellipsoidal *Konderia* (160 × 240 μm) are other characteristic components of the assemblage. Its colonial coccoidal forms are represented by *Ostiana* remains (8–10 and 23–24; hundreds μm) of different preservation state, including the cell decomposition stages "capsulata" and "punctata" (Hofmann, 1976), and also by trito pentaserial *Chlorogloeopsis* (12; 30–40 μm), racemose *?Palaeopleurocapsa* (16; 50–60 μm), and spheroid *Sphaerophycus* (4–8; 20–40 μm).

Predominant among abundant filamentous forms of the upper sublevel are sheaths and smooth-walled cuticular remains. The former are as wide as 1 to 90 μm and enclose numerous filaments and mats of *Eomycetopsis* (2–4 μm), *Leiotrichoides* (5–30 μm), *Asperatofilum* (up to 40 μm), *Taenitrichoides* (up to 90 μm), and ornamented *Siphonophycus* (25–40 μm) with a fine transverse structure. The mats can be as wide as 260 μm. Trichomes are represented by *Palaeolyngbya* (12–13; 20–22, and 40 μm) and by large forms resembling *Oscillatoria* (30–40 μ; cell length up to 50 μm). *Trachytrichoides* (20–22 μm; cell length more than 40 μm) is the sole trichomelike form. Microfossils of a complex structure include the conical membranes (70 μm) also present at the fourth level and cancellate *Fabiformis* (55 × 530 to 70 × 730 μm) of irregular ellipsoidal shape. Peculiar remains of a dumbbell-like shape are close in size to the latter; they have a little neck and enclose, in their polar areas, the small bodies (70 μm) with inner inclusions. These forms are similar to small *Pseudotawuia* described as remains of unclear taxonomic position from the Upper Riphean Svanbergfiellet Formation of Spitsbergen (Butterfield *et al.*, 1994). Two other morphotypes of the assemblage are (1) the linear aggregates of spheroid bodies (37–40 μm) lacking their terminal structures and consisting of subtriangular elements that enclose the abundant dark inclusions, and (2) solitary clavate forms (up to 150 μm) ornamented with rare conical processes in their wider parts.

At both sublevels, the microfossils are concentrated in the dark gray to black shales containing the abundant organic detritus and enclosing rare laminae of quartz–carbonate siltstones with inner graded bedding. These deposits accumulated in the proximal zone of an open

Plate II. Microfossils from the Upper Riphean Miroedikha Formation of the Turukhansk Uplift: (1, 2, 4, 8, 10, 11, 16, 18) *Oscillatoriopsis* spp., no. 2678-4389 (1), no. 2678-4388 (2), no. 2678-4683 (4), no. 2678-4299 (8), no. 2678-4386 (10), no. 2678-4311 (11), no. 2678-4298 (16), no. 2678-4313 (18); (3) convolute sheath of *Leiotrichoides typicus* Herm., emend. Herm., no. 4571-130; (5) *Filiconstrictosus majusculus* Schopf, no. 2678-4684; (6) *Cephalophytarion variabile* Schopf et Blacic, no. 2678-4211[4]; (7) *Calyptothrix perfecta* A. Weiss, no. 2678-4214; (9, 19) *Cephalophytarion turukhanicum* A. Weiss, no. 2678-421 (9), no. 2678-4297 (19); (13–15) trichomes with intercalary, terminal, or sectional enlarged cells (spores ?, heterocysts ?) similar in morphology to recent nostocalean cyanobacteria, no. 2678-4229 (13), no. 2678-4239 (14), no. 2678-4240 (15); (17) filamentous remain (sheath ?) with an irregular pseudocellular structure originated in the course of posthumous destruction, no. 2678-430; (20) fusiform C-shaped aggregate of thin filaments (sheaths) ex gr. *Eomicrocoleus crassus* Horod. et Donald. with closed narrowing terminations, no. 2678-4385; (21) *Asperatofilum experatus* (Herm.), no. 4075-200; (22) *Circumvaginalis elongatus* Serg., no. 907-162; (23, 24, 28) *Palaeolyngbya sphaerocephala* Herm. et Pyl., trichomes enclosed in the wide dense sheaths, no. 4075-110 (23), no. 2578-6845 (24), specimen illustrating the trichome decomposition into hormogonia, no. 4075-100 (28). (25, 27) *P. catenata* Herm., no. 907-905 (25), no. 907-799 (27); (29) *Botuobia* sp., trichome with shortened cells, no. 2678-4296.

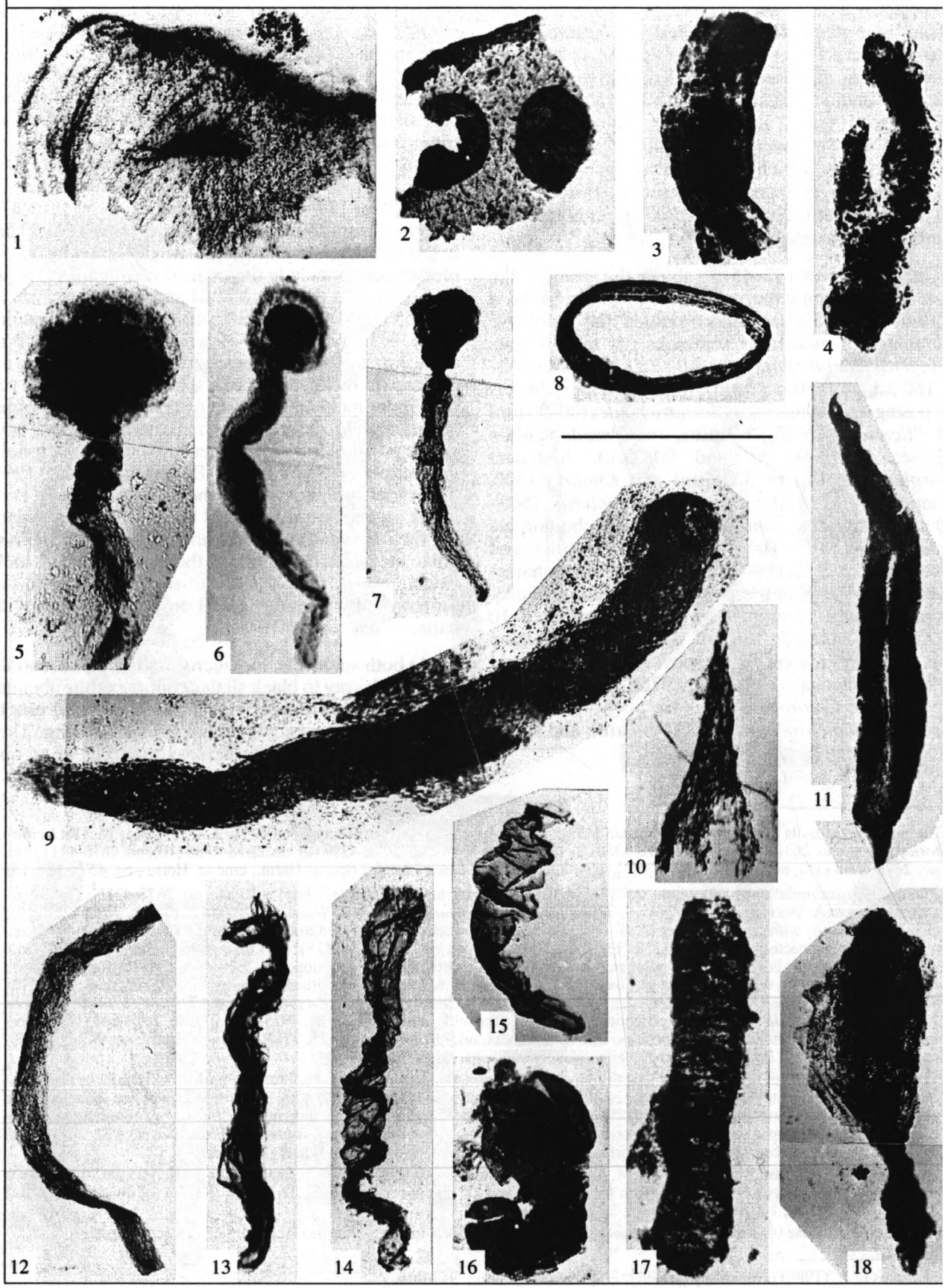
The occurrence levels of microfossils: (4, 5) third level, the second member of the lower subformation (Miroedikha River); (1–3, 6–20, 29) fourth level (ditto, excluding specimen 3 from the Bol'shaya Shorikha locality); (26, 27) fifth level, the first member of the upper subformation (Enisei River); (21, 23, 28) lower sublevel of the sixth level, the first member of the upper subformation (Miroedikha River); (25) upper sublevel of the sixth level, the first member of the upper subformation (Enisei River); (22, 24) seventh level, the second member of the upper subformation at the Miroedikha River(24) and Enisei (22) rivers.

Scale bar size: 40 μm for images 1, 2, 10, 11, 13, 16, 18; 130 μm for image 3; 60 μm for images 4, 5, 15, 25; 65 μm for images 6 and 17; 50 μm for images 7–9 and 14; 70 μm for images 12 and 27; 80 μm for images 19, 22, 26, 29; 210 μm for image 20; 160 μm for images 21 and 23; 100 μm for image 24; 260 μm for image 28.

All specimens whose images are presented are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

[4] Yankauskas *et al.* (1989) erroneously quoted this specimen under no. 944-946.

Plate III



shelf periodically affected by turbidity influx of the fine-grained carbonate and siliciclastic material. The hidden erosion surfaces in the middle part of this interval point to the influence of weak contour currents and, likely, of the storm waves.

Seventh level (0–1.5 m above the base of the second member of the upper subformation at the Miroedikha River and 0–2 m above it at the Enisei River) bears a peculiar assemblage, whose most characteristic microfossils are the monstrous acritarchs (120–170 μm), aggregates of colonial coccoidal *Coniunctiophycus* (10–14; 120 μm, and 30; 180 μm), and filaments of *Palaeolyngbya sphaerocephala* (40 μm). The assemblage also includes *Leiosphaeridia* (up to 300 μm), *Chuaria* (up to 560 μm), *Prolatoforma* (200–220 μm), *Trachyhystrichosphaera* (180–200 μm), envelopes resembling *Navifusa* (120–200 μm), and dense dark *Cerebrosphaera* (150 μm). The shapeless thalluses of *Ostiana* (20; hundreds μm), small *Polysphaeroides* (1–2; 20 μm), *Eogloeocapsa* (4–6; 48–50 μm), and spheroid fragments of aggregated *Coniunctiophycus* (10–14; 150 μm) and *Sphaerophycus* (9–13; 60 μm) are predominant among the colonial coccoidal microfossils.

Smooth-walled and ornamented sheaths together with tubercular cuticles(?) are most abundant and diverse among the filamentous forms of the seventh level. The first group includes the isolated *Asperatofilum* (50 μm), *Eomycetopsis* (2–4 μm), tightly woven mats of *Leiotrichoides* (5–16 μm), multiserial *Polytrichoides* (2–4 μm), and *Circumvaginalis* sheaths (10–13 μm). Small forms close in morphology to *Siphonophycus* (10–16 μm) and *Rugosoopsis* (35–40 μm) together with two specimens of dense remains (40 and 100 μm), complex in their structure and representing new taxa, are attributed to the second group. In addition to transitory trichomes, the group of septate forms includes the filaments of *Palaeolyngbya* (10, 12–13, and 15–17 μm), including *P. sphaerocephala* (40 μm), and single fragments of large trichomelike structures (50 μm) consisting of dark rectangular cells (10–12 × 50 μm). Complex forms of microfossils are represented by spiral *Elatera* (100 μm and larger), conical membranes (60–70 × 134 μm), *Fabiformis* (180 μm) with an unusual longitudinal structure, *Pseudodendron* thalluses (25–50 μm), and by very large elongated envelopes (up to 1000 μm) previously unknown and presumably having the multichamber structure.

Microfossils of the seventh level occur in the dark to greenish gray shales enclosing numerous structureless films of organic matter. These sediments are intercalated with limestone beds whose abundance grows upward the section, where the fine-grained clayey limestones are replaced by calcarenites and calcirudites sometimes filling the erosion channels. These deposits accumulated in the relatively deep distal shelf zone below the basis of wave activity, when the basin experienced some shoaling.

Eighth level (2–4 m above the base of the second member of the upper subformation at the Miroedikha River and 3–4 m above it at the Enisei River) contains the abundant acritarchs *Valeria* (230–250 μm), *Simia* (140–350 μm), *Nucellosphaeridium* (up to 200 μm), *Ostiana* thalluses (12–19 μm), empty colonies of *Coneosphaera* (10–12, 30–32; 125, and 250 μm), *Polytrichoides* sheaths (4–6 μm), and the intergrowth structures of fungi and/or algae *Caudosphaera* (50–180 μm, with processes up to 200 μm long). These forms associate with the following taxa: *Leiosphaeridia* (up to 300 μm); *Chuaria* (600–630 μm); *Ch. globosa* (350 μm); platty *Satka* (up to 200 μm); envelopes resembling *Navifusa* (160–400 μm); single monstrous acritarchs (up to 350 μm); *Trachyhystrichosphaera* with low widened processes (160 μm); dark tubercular envelopes (100 μm); and new *Valeria* morphotypes (200 × 260 μm)

Plate III. Microfossils from the Upper Riphean Miroedikha Formation of the Turukhansk Uplift: (1) *Elatera binata* Herm., tubercular cuticular(?) forms with a fine transverse striation producing rhomboid patterns in superimposed specimens, no. 2578-6846; (2) fragment of a wide thallus of problematic Metaphyta with large dark bodies, no. 2578-6450; (3) *Eosolena anisocyta* Herm., articulate forms divided into chambers and enclosing aggregates of rounded bodies (cells) in their central parts, no. 2678-4321; (4) branching thallus displaying a kind of cellular structure in the wider basal part, no. 2678-4230; (5–7) *Caudosphaera expansa* Herm. et Timof., no. 2578-6401 (5), no. 2578-6100 (6), no. 2578-6410 (7); (8) ringlike fragment of an organism whose morphology is unknown, no. 2578-4300; (9) remain of a large helminthoid organism with the transverse and/or tangential ornamentation, no. 2678-4220; (10) acute-conical filmy form resembling the Early Cambrian microfossils *Ceratophyton*, no. 2578-6300; (11) large fusiform remain ex gr. *Pellicularia* displaying the longitudinal rima and cellular structure in its acute terminations, no. 2578-6399; (12) *Pseudodendron* sp., thallus with the longitudinal structure exemplifying ancient scytonemean cyanobacteria, no. 2578-6498; (13–15) thin filmy thalluses (possibly remains of siphonocladous green algae ex gr. *Proterocladus*), no. 2578-6499 (13), no. 4571-200 (14), no. 907-180 (15); (16) elongated sublinear aggregate of cells with budlike termination, no. 2678-4210; (17) very large elongated-ellipsoidal envelope with suspicious multichamber structure, no. 4084-600; (18) clavate remain with low conical processes in the wider part, no. 4085-100.

Scale bar size: 100 μm for images 1, 5, 9; 500 μm for images 2 and 17; 250 μm for image 3; 170 μm for image 4; 200 μm for images 6, 7, 14, 15; 70 μm for image 8; 170 μm for images 10, 13, 16; 800 μm for images 11; 330 μm for image 12; 250 μm for image 18.

The occurrence levels of microfossils: (3, 4, 9, 14, 16) fourth level, the second member of the lower subformation at the Miroedikha (3, 4, 9, 16) and Bol'shaya Shorikha (14) rivers; (2, 5–7, 10–13) lower sublevel of the sixth level, the first member of the upper subformation (Miroedikha River); (15) upper sublevel of the sixth level, the first member of the upper subformation (Enisei River); (1, 17) seventh level, the second member of the upper subformation (Miroedikha River); (8, 18) eighth level, the second member of the upper subformation (Miroedikha River).

All specimens whose images are presented are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

unusual in their pyriform shape and hypertrophied lophostriate patterns. All listed taxa coexist with abundant, but not diverse, colonial coccoidal forms and trichomes. The former occur as dense spheroid aggregates, the fragments of *Coniunctiophycus* colonies (10; 40–60 μm, and 14–30; 70 μm), while the latter are represented by *Cyanonema* with elongated cells (12 μm, cells up to 15 μm long), *Palaeolyngbya* (9, 12, and 40 μm), trichomelike *Trachytrichoides* (12 μm), and also by remains of doubtful bangiophytes (10–12 μm). Remarkable sheaths of the level are also not very diverse. These are *Asperatofilum* (40 and 52–54 μm), *Taenitrichoides* (82–86 μm), *Archaeoellipsoides* (30 × 110 μm), and aggregates of thin filaments (1–2 μm) similar to *Eomycrocoleus*, but enclosed in a single fusiform envelope (40 × 300 μm) that is closed at both terminations. *Elatera* (fragments 100–120 μm wide), thalluses with a vague longitudinal structure (50–80 μm), and formerly unknown envelopes (26 × 120–140 μm) with numerous necks and open subspheroid ends are attributed to complex microfossils.

All the taxa are recovered from the relatively thin (20–30 cm) beds of greenish gray shales intercalated with thicker (up to 50 cm) beds of clastic limestones (calcarenites and calcirudites). In general, this level is similar to the previous one in sedimentation environments, but the observed, more frequent signs of storm wave activity suggest the further shoaling of the basin.

Ninth level (4.6–5.2 m above the base of the second member of the upper subformation at the Enisei River) is characterized only by transitory taxa of the first category described above. They are detected in the rather thin (15–17 cm) interbeds of greenish gray shale enclosed in clastic limestones dominant in this part of the section. These deposits accumulated in the upper subtidal zone under the persistent influence of wave activity. This is evident from beds (up to 1 m) of oolitic grapestones present in the section and characteristic of the subtidal carbonate shoals, and also from small redeposited stromatolitic bioherms at the base of these beds.

Tenth level (12 m above the base of the second member of the upper subformation at the Enisei River) encloses a very impoverished assemblage of transitory taxa similar to those of the ninth level. The microfossils are concentrated here in thin (2–5 cm) lenticular interbeds of dark gray shales intercalated with clastic and stromatolitic limestones. These sediments accumulated in the upper subtidal zone frequently affected by the storm wave activity (coarse-grained flakestones) that disturbed the general wave regime of a shoal (doloarenites) located behind the stromatolitic barrier system. The shale interbeds bearing microfossils accumulated in local settings protected by this system from the wave influence.

The higher horizons of the Miroedikha Formation contain only rare acritarchs, some colonial coccoidal and filamentous forms of a low preservation degree, and some interbeds enriched in organic matter. This microphytological "defect" of the uppermost part of the formation apparently is a result of certain factors presumably related to peculiar features of the late Miroedikha biota still poorly understood, and to fossilization regime in the basin of that time. The most obvious of the negative factors could be related to the slowing down rate of clayey sedimentation and to parallel acceleration of the posthumous microbial decomposition of microorganisms.

FACIAL–ECOLOGICAL DISTRIBUTION OF THE MIROEDIKHA MICROFOSSILS

As noted above, the suggested model of the Miroedikha basin evolution (Petrov and Semikhatov, 1997) elucidate the sedimentation environments for each of ten distinguished microphytological levels. Accordingly, this offers a possibility to reveal the distribution patterns of different assemblages of microorganisms along the bathymetric profile of the past basin, to understand some peculiar features of microbiota during the transgressive and regressive stages of the basin evolution (including the stable period of the latter), and to demonstrate the facial–ecological behavior of some indicative morphotypes.

In our analysis of the paleoecological structure of the Miroedikha microbiota, we quite conventionally subdivided all taxa into eurybionts, mesobionts, and stenobionts. The first group includes the taxa, which have been encountered in sediments of all four zones of the basin: the upper and lower subtidal zones, and the proximal and distal shelf. Taxa of second group inhabited two or three zones not obligatory conjoined with one another, and the last organisms were characteristic of a single zone only.

The most distinct patterns of the lateral distribution of the Miroedikha microbiota are as follows: microfossils from shallow settings of the upper subtidal zone (assemblages of the first, second, ninth, and tenth levels) are the least diverse and abundant as compared to others; abundance and diversity of microfossils grow in depositional settings of the lower subtidal zone (assemblages of the third, seventh, and ninth levels along with that of the fourth level at the Bol'shaya Shorikha River), and especially in the moderately deep zone of the proximal shelf (assemblages of the fourth and sixth levels at the Miroedikha and Enisei rivers); quick and selective decline in abundance and diversity of certain taxa is characteristic of deeper and distal settings of sedimentation (assemblage of the fifth level).

As we will see below, these features are distinguished at both the transgressive and regressive stages of the basin evolution and depict the most general trend of the lateral distribution of microbiota. In addition, there is a distinct negative correlation between the extent of facial–ecological distribution of microfossils and their abundance, on the one hand, and between the first parameter and diversity of taxa, on the other. Eury-

bionts occur in all samples from all zones and are as abundant as hundreds and thousands of specimens, but they belong to a limited number of taxa, whereas stenobionts from restricted distribution areas are represented by not abundant, but very numerous taxa. In both characteristics, mesobionts occupy an intermediate position between steno- and eurybionts, although they demonstrate sometimes the local peaks of abundance inside their distribution areas. We should also note that the major part of stenobionts representing the most characteristic components of microbiota consist of certain subassemblages of their morphotypes, presumably similar in the biological aspects. These subassemblages predominantly include endemic taxa along with genera and/or species not yet described in publications.

Eurybionts of the Miroedikha microbiota are represented by simple acritarchs, colonial coccoidal, and filamentous forms of six genera only: *Leiosphaeridia*, *Myxococcoides*, *Eomycetopsis*, *Leiotrichoides*, *Oscillatoriopsis*, and *Polytrichoides*. Representatives of the last genus are predominantly typical of deposits accumulated in the moderately deep settings. They were very rare in both the deep distal shelf and shoal settings, and some of them migrated upward up to the boundary settings between the upper and lower subtidal zones at the transgressive stage only. All eurybiont genera represent the primitive planktonic eukaryotes and both planktonic and benthic cyanobacteria. All identified *Leiosphaeridia* can be quite positively attributed to the former, whereas *Eomycetopsis* and *Leiotrichoides* forms that constructed the extended mats very characteristic of subtidal settings of the Miroedikha basin belong undoubtedly to the benthic prokaryotes. The discrete specimens of their sheaths together with *Polytrichoides* sheaths, *Oscillatoriopsis* trichomes, and coccoidal *Myxococcoides* could be equally the components of planktonic and benthic coenoses. In general, the dominant eurybionts of the Miroedikha basin correspond to the well-known transitory taxa of the late Precambrian.

The most remarkable stenobiont taxa represent two subassemblages that were localized during the transgressive and regressive stages in two discrete biotopes: in the lower subtidal zone and in the proximal shelf. In the first zone, the characteristic archetype is exemplified by wide tubercular, presumably cuticular forms: at the transgressive stage (fourth level at the Bol'shaya Shorikha River), they included the smooth-walled *Taenitrichoides* and *Plicatidium*, whereas at the regressive stage, the larger forms of the same genera coexisted with *Rugosoopsis* and *Pseudotawuia*, and also with abundant, but still unnamed tubercular morphotypes displaying the transverse ornamentation, and with very large (greater than 1 mm) elongated envelopes likely divided into chambers.

The subassemblage of the proximal shelf is more diverse in morphological and taxonomic aspects. At the transgressive and regressive stages, it included the linear aggregates of cells with the budlike terminations, *Trachytrichoides*, *Eosolena*, and *Fabiformis*. During the transgression, they coexisted with trichomelike *Arthrosiphon* and *Cephalonyx*, as well as with ornamented *Rectia* and diverse trichomes, the heterocyte type of the latter inclusive. At the regressive stage, this subassemblage also incorporated very large thalluses of suspicious *Metaphyta*, a new form of large *Pellicularia*, and some morphotypes unknown from publications.

Morphological features and size ranges of most stenobionts from the Miroedikha Formation (apart from trichomes and some trichomelike forms of the fourth level) indicate that they cannot be classed with prokaryotes. Localization of these microfossils in the narrow shelf zone suggests that they were benthic organisms, and peculiar morphology of most specimens was interpreted as evidence of the possible heterotrophic nature of some of them (German, 1990). We would like to stress that the isolated communities of such microorganisms, which were concentrated in particular biotopes, are unknown in the North Eurasian deposits older than the Miroedikha Formation.

Stenobionts of the formation that cannot be considered as components of the indicated subassemblages are rare and diverse in morphology. At the regressive stage, these were the following morphotypes: cancellate *Spumosina* (fifth level), *Cerebrosphaera* (seventh level), and acritarchs with the hypertrophied lophostriation (eighth level). At the transgressive stage, only *Arctacellularia* and *Sphaerocongregus* (third level) represented this type of microfossil. Despite the evident heterogeneity of these forms (planktonic to benthic eu- and prokaryotes), they all were concentrated in the relatively shallow settings of the subtidal zone.

In addition to the listed taxa, the studied microbiota includes some other genera (*Palaeopleurocapsa*, *Circumvaginalis*, *Archaeotrichion*, *Obruchevella*, and some others) whose representatives were also encountered as single specimens in our rock samples of the restricted facial affinity. However, in other Precambrian biotas, they are known as occurring in a wider spectrum of habitat areas (Hofmann, 1976; Knoll and Golubic, 1979; Sergeev, 1992), and consequently these forms have not been attributed to stenobionts.

Mesobionts are very abundant in the Miroedikha microbiota and represent more than 50% of all identified genera. The following forms most frequently occur amid them: medium-sized *Chuaria* (*Ch. globosa* inclusive), immarginate *Trachyhystrichospaera*, small *Nucellosphaeridium*, *Simia*, envelopes of the *Navifusa* type, all monstrous acritarchs, cenobial aggregates of *Leiosphaeridia*, *L. bicrura* Jank., *L. ternata* (Tim.), *Sphaerophycus*, *Palaeolyngbya* (*P. sphaerocephala* inclusive), *Caudosphaera*, and most narrow thalluses of *Pseudodendron*. These taxa inhabited all shelf areas of the Miroedikha basin except its upper subtidal zone. More limited distribution areas were characteristic of largest *Chuaria*, *Nucellosphaeridium*, *Pterospermopsi-*

morpha, and most complex unclassified forms, all generally concentrated in the proximal shelf. *Coniunctiophycus*, problematic bangiophyte remains, and *Elatera* usually were abundant in the lower subtidal zone. Monstrous acritarchs and microfossils complicated in structure are the mesobionts concentrated in sediments of the regressive stage. The facial control over others was quite similar at both the regressive and transgressive stages. This inference equally concerns the typical dominant mesobionts of the Miroedikha shelf, such as *Chuaria* of many size generations, *Ostiana*, *Asperatofilum*, *Taenitrichoides*, *Cephalophytarion*, *Filiconstrictosus*, *Palaeolyngbya*, *Germinosphaera*, *Pseudodendron*, marginate *Trachyhystrichosphaera*, *Prolatoforma*, *Simia*, *Valeria*, *Cucumiforma*, *Konderia*, *Polysphaeroides*, *Coneoshaera*, *Archaeoellipsoides*, *Brevitrichoides*, *Glomovertella*, and others. They all were rather homogeneously dispersed in the lower subtidal zone and proximal shelf, and only some local settings concentrated these forms.

CONCLUSION

The considered data show that each of four principal zones of the Miroedikha shelf is characterized by its own assemblages of microfossils. The upper subtidal settings were inhabited by eurybionts limited in abundance. In deeper settings located seaward, these forms coexisted with various meso- and stenobionts. As a result, communities of microfossils in the lower subtidal zone and proximal shelf were most abundant and diverse, and included almost all morpho- and ecotypes identified in the Miroedikha microbiota. Communities of the distal shelf, the least diverse in general, were especially depleted in meso- and stenobionts, and the latter were represented here only by one genus of problematic acritarchs. This type of the lateral distribution of microfossils was characteristic of both the transgressive and regressive stages of the basin evolution. Accordingly, the ecological structure of microbiota remained in a steady state all the time despite a considerable spatial migration of facial zones.

Studying some other Riphean microbiotas from Siberia, we demonstrated earlier that principal factors responsible for the distribution patterns under consideration are the bathymetric control, hydrodynamic activity of water masses, and the state of biota at the transgressive, regressive, and relatively stable stages of the basin evolution. We also distinguished four facial–ecological biotic groupings different in composition of microfossils: shallow-water, transitional, medium-deep or "optimal," and deep-water (Veis and Petrov, 1994; Petrov and Veis, 1995). Contribution of these factors in the lateral distribution of the Miroedikha microfossils is considered below (Fig. 2).

Changes in the basin depth with time affected the dynamics of the complete recolonization, whereas lateral variations of this factor at any given moment controlled the general lateral structure of the autochthonous Miroedikha biota, like in other cases studied previously. The distribution areas of mesobionts, and, especially, of stenobionts were strictly limited by particular segments of the bathymetric profile, whereas eurybionts inhabited all zones of the basin being slightly variable in composition by passing from one zone to another. Stenobionts, excluding the deep-water *Spumosina*, inhabited only the moderately deep zone, but a considerable part of mesobionts were able to migrate outside this zone. For instance, *Glomovertella* and *Polytrichoides* migrated in the upper subtidal settings, and the same *Polytrichoides* in association with *Chuaria*, *Trachyhystrichosphaera*, and some other taxa penetrated into the distal shelf, where they composed together with eurybionts the characteristic impoverished communities.

Thus, every facial–ecological zone of the Miroedikha and other Riphean basins hosted its own continuum of organisms different in biological and ecological aspects. Biota of all shelf zones included assemblages of simple phototrophic eu- and prokaryotic eurybionts. By passing to deeper settings, their communities incorporated first the more evolved pro- and eukaryotic photo- to heterotrophic mesobionts, and then, stenobionts exemplifying the highly organized members of biota. In the deepest zone, the biotic continuum became deprived of both additional assemblages. In other words, the basin depth was the principal abiotic factor that controlled the biota composition from the time of the basin colonization by the most tolerant eurybionts to the later appearance of mesobionts and stenobionts, the last of which were most sensitive to environmental factors and concentrated in particular zones.

Hydrodynamic fluctuations in the proximal shelf zone also controlled the Miroedikha biota composition. Eurybionts reacted to this factor less dramatically than others, and their abundance declined insignificantly in response to growing hydrodynamic activity. For meso- and stenobionts, this factor was more influential and restricted the expansion of pro- and eukaryotic benthos (*Pseudodendron*, *Sphaerocongregus*, problematic bangiophytes, and some complex forms) into the upper subtidal zone. At the same time, it was favorable for migration of some remarkable planktonic taxa (very large *Chuaria*, *Valeria*, *Trachyhystrichosphaera*, *Cucumiforma*, *Simia*, *Coneosphaera*, and others) into the relatively shallow settings. Finally, we assume that the distribution areas of subassemblages of the wide tubercular (cuticular ?) stenobionts characteristic of the quiet settings in the lower subtidal zone of the inner shelf were under strong control of the hydrodynamic factor.

In the case of the Miroedikha microbiota, the influence of transgressive and regressive trends on the taxonomic composition of microfossil communities is evident only from a certain localization of complex morphotypes of meso- and stenobionts, especially of those, which presumably are products of the biological evolu-

tion. In this particular case, we should emphasize three important points:

(1) Many complex forms and almost all evolutionary innovations are detected at the regressive stage in a wide series of settings of the lower subtidal zone and proximal shelf. Precisely here, we encountered new forms of *Pellicularia* and *Valeria*, almost all monstrous acritarchs, *Cerebrosphaera*, *Elatera*, ornamented cuticular(?) morphotypes, and many forms previously unknown. (2) At the transgressive stage, these and similar elements of biota were considerably less diverse and abundant. They appeared for a short period just at the end of the stage and in the deepest settings of the proximal shelf. (3) Apart from *Arthrosiphon*, *Cephalonyx*, *Eosolena*, *Polysphaeroides filiformis*, and some unclassified complex morphotypes, all remarkable meso- and stenobionts of the transgressive stage were represented by transitory taxa.

Thus, the transgressive and regressive stages controlled solely the composition and distribution of the highly organized meso- and stenobionts. This resulted in the growing abundance and diversity, and in the lateral expansion of complex microfossils at the regressive stage only. In our forthcoming paper, we will consider this phenomenon in detail to show that it is a result of both the abiotic changes and autogenetic biota evolution.

Thus, the revealed general lateral structure of the Miroedikha microbiota divisible into four taxonomically different and spatially interrelated groupings (shallow-water, transitional, "optimal," and deep-water) was characteristic not only of the transgressive, but of regressive stage as well. Simultaneously, it had some distinct features during the particular periods of time. At the both stages, the general paleoecological features of this biota were as follows: (1) the shallow-water groupings inhabited the upper subtidal zone only, and their taxa were almost identical in the time periods of the first, second, ninth, and tenth levels; (2) characteristic representatives of the "optimal" grouping were concentrated in the deepest settings of the lower subtidal zone and open proximal shelf (assemblages of the fourth, sixth, and, to some extent, of the sevens level); (3) transition from the "optimal" to deep-water grouping at the transgressive stage was marked by a drastic decrease in abundance and diversity of microfossils, but both of these parameters were completely restored during the backward transition from the deep-water to "optimal" grouping at the regressive stage (assemblages of fourth–fifth and fifth–sixth levels, respectively).

Therefore, the principal paleoecological features of microorganism communities from the Miroedikha basin are similar to those established for other Riphean basins of Siberia (Veis and Petrov, 1994; Petrov and Veis, 1995). At the same time, there are some peculiar features of the Miroedikha biota distribution. First of all, an impoverished transitional grouping that inhabited the lower subtidal zone at the transgressive stage (assemblage of the third level) is missing in deposits of the regressive stage. During the last stage, the assemblage of the lower subtidal zone of comparable bathymetric position (microfossils from the seventh level) was more diverse in taxonomic aspect and rather close to the "optimal" grouping. Another peculiar feature of the Miroedikha biota is exemplified by an unusual composition of its deep-water grouping (assemblage of the fifth level). Surprisingly, it turned out to be similar to the transitional grouping of the third level despite a some difference in quantitative proportions of taxa. However, this deep-water grouping includes the cancellate *Spumosina* envelopes, interlaced narrow thalluses of *Pseudodendron*, and intergrowth structures of fungi and/or algae *Caudosphaera*, all untypical of true transitional groupings of the Miroedikha and other Riphean basins of Siberia.

These specific features of the deep-water assemblage in question could be the results of taphonomic and facial–ecological factors. The possible contribution of taphonomic factors is suspected on the basis of two circumstances. First, the deep-water clayey rocks of the fifth level enclose the peculiar microbeds enriched in the decomposed organic matter, suggesting the former existence of coenoses of abundant destructive benthic organisms. Second, this matter is concentrated in the gray and green rock varieties only, while in the intercalated red beds it is missing. As one more indirect evidence, we should mention that the Precambrian deposits, whose deep-water origin is well substantiated, are extremely poor in remains of microorganisms, because the fossilization potential of the latter apparently was very low in general (Oehler, 1977; Vidal and Nystuen, 1990). The influence of facial–ecological factors on the composition of microfossils from the fifth level could be understood as a consequence of the relatively shallow depth of the distal Miroedikha shelf that was presumably less than hundred meters deep (Petrov and Semikhatov, 1997). These environments caused the reduction of the "optimal" grouping, but were insufficient for the mature development of the deep-water organic community. As a result, the latter was replaced by relic forms of the "optimal" grouping, i.e., by eurybionts, by most tolerant mesobionts, and by rare deep-water stenobionts in the strict sense.

Thus, despite some peculiar features, the facial-ecological distribution patterns of the Miroedikha microbiota are in a good agreement with the general spatial arrangement of microorganisms that was established in the course of previous investigation of Riphean basins. These distribution patterns of microorganism communities were controlled by the bathymetric profile of a given basin, by its hydrodynamic activity, and by the state of biota during the transgressive and regressive stages in the basin evolution. Judging from all data available at present, the Precambrian biota was already significantly differentiated and consisted of qualitatively inhomogeneous components. As a result, shal-

low, transitional, and deep zones in the Riphean basins were populated by microorganisms contrast in their ecological parameters, but simple in general morphology and showing a very slow rate of biological transformations. In contrast, the distinguished "optimal" grouping (community) of microbiota that persistently occupied the same medium-deep ecological niches in the basins included not only simple and conservative, but also the complex forms of microorganisms. Precisely these forms were responsible for the characteristic morphological innovations attributable to the manifestation of macrobiological evolution that progressed against the background of constant changes in the microbiota composition, which were interrelated with evolutionary megacycles of the Riphean basins. Leaving aside all details, we may conclude that such an "optimal" grouping appeared in the Precambrian basins when their biological substratum and ecological niches became suitable for the evolutionary experiments of the ancient biota and were sufficiently safe to protect it from the total degradation in the case of negative results of these experiments. The detailed chronological analysis of changes in the Miroedikha microbiota will be published in the next our paper.

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Microfossils in Microphytolites of the Paleoproterozoic Gunflint Formation, Southern Canada

V. N. Sergeev, M. A. Semikhatov, and L. M. Mudrenko

Geological Institute of Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Abstract—Silicified microphytolites of the *Osagia* group from the Paleoproterozoic (Lower Proterozoic) Gunflint Formation of southern Canada (ca. 1.9 Ga) contain well preserved remnants of microorganisms represented by filamentous forms of genus *Gunflintia* (*G. minuta, G. grandis*), which compose accumulations (mats) along the *Osagia* lamination patterns, and by less frequent coccoidal *Huroniospora* sp. scattered inside the mats. The filamentous forms seem to be remnants of oscillatoriacean or nostocalean cyanobacteria, and the latter ones are those of unicellular chroococcalean cyanobacteria. The same microfossils were found in the semirounded flattened fragments of stromatolitic laminae, well rounded pellets, and some small columnar stromatolites occuring in the samples studied. Siliceous matrix of the samples contains the scarce planktonic microfossils only. The predominant representatives of genus *Gunflintia* inside *Osagia*, as well as their occurrence mode, support the earlier suggestion that filamentous hormogonian cyanobacteria took part in the formation of concentrically laminated microphytolites and composed dense mats on their surfaces. Microfossils from the *Osagia* group are taxonomically identical to those found earlier in silicified stromatolites of different morphology from the Gunflint Formation, but proportions of the present taxa are different.

Key words: *microphytolites, microfossils, cyanobacteria, oncolites, stromatolites, Paleoproterozoic, Lower Proterozoic, North America.*

Although remnants of microorganisms are rarely preserved in the Precambrian silicified microphytolites, the formation of these biolites is commonly considered to have been significantly affected by microbial associations. At present, occurrences of microfossils within the silicified Precambrian microphytolites are known only in a few localities. They are reported from the Paleoproterozoic (Lower Proterozoic) Gunflint Formation of North America and Frere Formation of Australia, and also from the Upper Riphean–Vendian Chichkan Formation of southern Kazakhstan (Barghoorn and Tyler, 1965; Walter *et al.*, 1976; Awramik and Barghoorn, 1977; Hofmann and Schopf, 1983; Sergeev and Mudrenko, 1997). Endolithic microorganisms are more frequently preserved inside the Precambrian microphytolites (Knoll *et al.*, 1986, 1989; Green *et al.*, 1988), but they penetrated into the biolites when the latter had been already lithified and therefore were unrelated to their formation. As for the Cambrian microphytolites, some of them enclose carbonate microfossils (Yakshin, 1978) owing to the appearance of algae with calcareous sheaths at that time (Luchinina, 1988; Voronova, 1976). However, considerable changes in the biosphere near the Precambrian–Cambrian boundary (Brasier, 1992; Schopf and Klein, 1992; Knoll *et al.*, 1993) made it impossible to use the Paleozoic material in modeling the formation of Precambrian microphytolites.

In this article, our attention is primarily focused on exceptionally well preserved silicified microfossils from microphytolites of the Paleoproterozoic Gunflint Formation of southern Canada, as well as on the morphology and taxonomy of these microphytolites. In addition, silicified structures of different morphology and genesis (fragments of stromatolitic laminae and pellets) containing silicified microfossils are also described, and compositions of the latter in different biolites are compared.

In the previous publications concerning remnants of the microorganisms from the Gunflint Formation (Barghoorn and Tyler, 1965; Awramik and Barghoorn, 1977), the emphasis was placed on the composition of microfossils, whereas microphytolites were inadequately described. They were only characterized by Hofmann (1969), who described oncolites together with stromatolites from the mentioned formation. The oncolites were identified as "Form G" (Hofmann, 1969, pp. 18 and 19) and assigned to the group *Osagia* Twenhofel, but no information about fossilized remnants of cyanobacterial associations was reported. However, the material from the Gunflint Formation orginally suggested a correlation between the morphology and stromatolites microstructures, on the one hand, and statistical peculiarities of the taxonomic composition of the enclosed microorganisms—possible stromatolite-builders, on the other (Awramik, 1976; Awramik and Semikhatov, 1979).

Therefore, a comparative analysis of taxonomical compositions of microfossils from the Gunflint and other Precambrian microphytolites different in morphology is of a special interest (Sergeev and Mudrenko, 1997). Such a comparison could be a first step in establishing a correlation between the morphological features of microphytolites, which were considered once as a promising tool for the Precambrian biostratigraphy (e.g., Zhuravleva, 1964), and the composition of enclosed microfossils.

GEOLOGICAL POSITION AND AGE CONSTRAINTS FOR THE SAMPLES STUDIED

The material used in this article was collected by Semikhatov in 1973, 10 km to the west of the Schriber Settlement (figure) at the northwestern shore of the Lake Superior, Ontario, Canada (outcrop no. 78418 according to Hofmann, 1969 and known as the Winston Point locality). This material is stored at the Geological Institute of Russian Academy of Sciences, collection no. 4313. The samples are collected from basal beds of the Gunflint Iron Formation of the Animikie Group. The latter lies with a sharp unconformity on the Archean granite–greenstone complex and represents the relatively thin (1.5–2.0 km), mildly altered (up to the greenschist facies) and slightly deformed predominantly sedimentary succession, which was accumulated in the northern margin of the Paleoproterozoic basin located in the vicinity of the Lake Superior and the eastern Lake Michigan shore (Douglas, 1972; Sims *et al.*, 1981; Morey and Van Schmus, 1988). In the southern part of the basin, there was an accumulation of Paleoproterozoic strata of predominantly volcanogenic composition, which include the pre-Animikie deposits pinching out to the north.

In the northwestern shore of the Lake Superior, the Gunflint Formation is separated from the Archean basement by a thin (up to 2 or 3 m) discontinuous horizon of quartzites and conglomerates of the Kakabekia Formation. The Gunflint Formation is represented there by the intercalated shales, siliceous and ferruginous rocks (takonites) locally enclosing stromatolites and microphytolites along with graywacke, and rare volcanic and dolomite beds. In the Winston Point section, the Gunflint Formation shows a basal variegated siliceous–ferruginous member 1.2–1.5 m thick and containing stromatolitic buildups with associated microphytolites (the so–called "lower algal member"). Upsection, there are shales, whose apparent thickness is 10–12 m, and, in the southern areas, other overlying members of the formation.

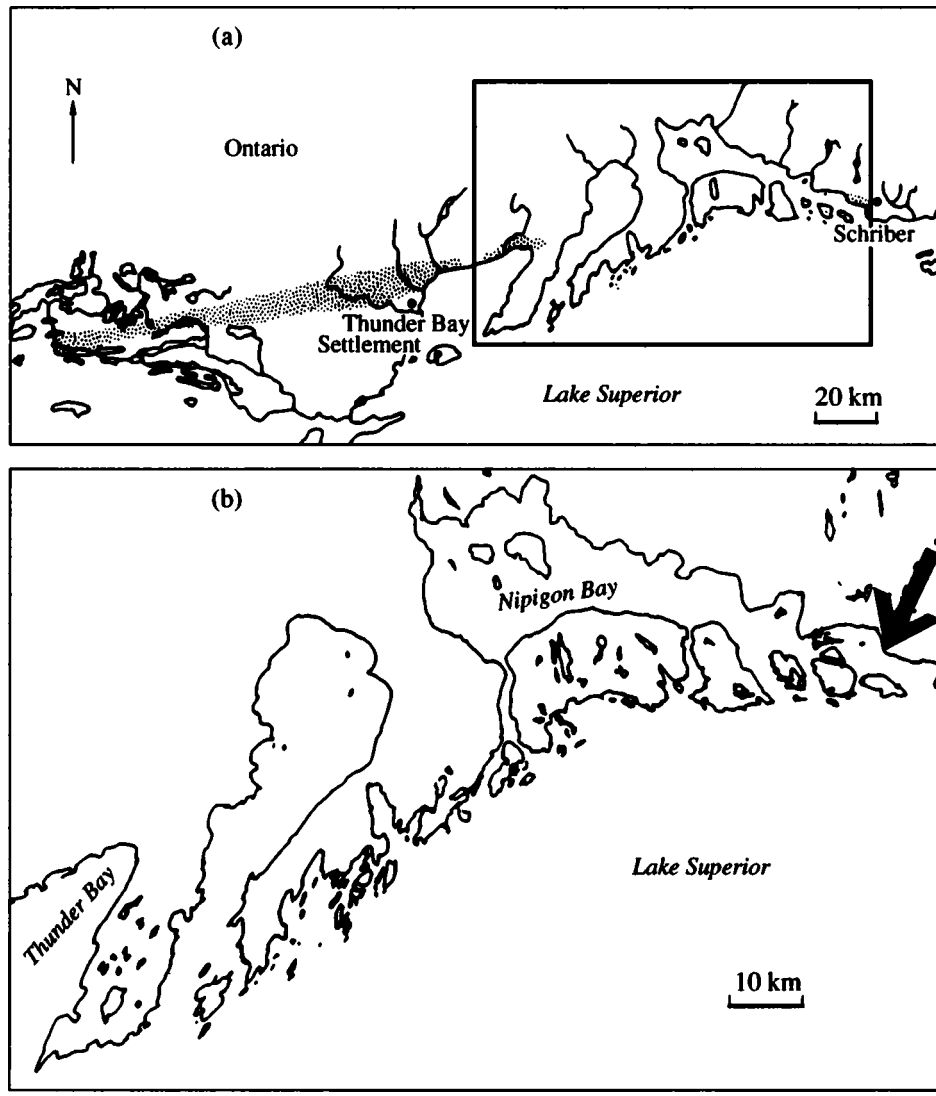
The stromatolites constitute isolated or connected bioherms varying from 10–12 cm to 1.5 m in diameter and from 15 to 45 cm in height. Hofmann (1969) emphasized their specific features, such as the small size, fine lamination, presence of well preserved silicified microfossils (mainly in the dark-colored rock varieties), and close association with siliceous spheroids (oolites and oncolites). The stromatolites are variable in morphology and mostly represented by small, frequently branching columnar forms (Hofmann, 1969; Cloud and Semikhatov, 1969; Awramik and Semikhatov, 1979). Our collection contains small irregularly branching forms with discontinuous walls, which are similar with the buildups of "Form B" described by Hofmann (1969). However, the available material (several small petrographic thin sections) is unsuitable for a more accurate comparison. The columns do not exceed 8–10 mm in diameter, and the intercolumnar space is from a few parts a centimeter to 2 or 3 cm wide. In the intercolumnar space, there are usually situated the white, gray, greenish, red and black silicified oncolites, elongated flat fragments (flakes) of stromatolitic laminae, and rounded unlaminated siliceous clasts (pellets) enclosed in the microgranular cherty matrix. The silicified oncolites with microfossils described in this article mostly occur in the intercolumnar matter, but some of them were found inside the stromatolitic laminae.

The Animikie Group is usually considered to be about 1900–2000 Ma old (Hofmann and Schopf, 1983; Schopf and Klein, 1992 and references therein). The U–Pb age of volcanogenic zircons from the analogues of the lower part of the Animikie Group in the southern shore of the Lake Superior is 1910 ± 10 Ma. The whole group was involved in the Penokean orogeny peaked at 1840–1850 Ma ago (Morey and Van Schmis, 1988; Hofmann, 1988).

MICROPHYTOLITES AND ENCLOSED MICROFOSSILS OF THE GUNFLINT FORMATION

The microphytolites and microfossils of the Gunflint Formation were studied in the standard petrographic thin-sections in translucent light. The size of the Gunflint microfossils makes it possible to obtain representative results on their size and morphology by studying the 30-μm-thick thin-sections. As for the microphytolites considerably larger in size (up to 0.7 mm in diameter), they are found to be represented in the slides by their particular sections. However, the study and diagnosis of Precambrian microphytolites are based on their sections in petrographic slides (Zhuravleva, 1964), although there are some difficulties in comparing the taxa (Mudrenko, 1991, 1992). In our case, the diagnosis was complicated by the silicified form of microphytolites, whereas their valid taxa were identified from carbonate forms essentially different in microstructure.

The studied oncolites are represented by concentrically laminated nodules 0.3–0.7 mm in diameter (Plate I, images 1, 8, and 10; Plate II, image 3). Their structure is determined by alternating and well persistent dark- and light-colored laminae varying in thickness from 5–8 to 30–35 μm. The laminae are dark brown, dark yellow, and light yellow or light gray in color. They frequently contain some amount of microfossils of vari-



Schematic map of the northwestern shore of the Lake Superior and the sampling place of the studied silicified microphytolites.

able but commonly very good preservation degree. These microphytolites belong to the group (formal genus) *Osagia* Twenhofel and bear a certain similarity to some formal species of this group (see below), though they probably represent its new forms. It is difficult to identify them more precisely because of silicification. It should be noted that the so–called "synthetic nodules" of *Osagia* (individual nodules covered by a single envelope) are extremely rare in our collection, but these forms were described by Hofmann (1969, pp. 9, 60, and 61).

The microfossils found in the microphytolites are confined generally to the dark laminae being predominantly represented by filamentous forms of genus *Gunflintia* Barghoorn (Plate I, images 2, 3, and 9; Plate II, images 4 and 5). Among representatives of this genus, two species can be distinguished: frequent trichomes 1–2 μm in diameter identified as *G. minuta* Barghoorn and sparse relatively big trichomes (3–5 μm) identified as *G. grandis* Barghoorn. By complicatedly interlacing, these filamentous microfossils form dense colonies (mats) up to few tens of microns thick; the filaments are oriented mostly perpendicular to the laminae. The mats are confined to the dark oncolitic laminae; they can be their major constituents imparting them a character of a "cancellate tissue" (Plate I, images 2 and 10; Plate II, image 4). The filaments of *G. minuta* and *G. grandis* frequently tend to be worse preserved, less closely packed, and gradually disappear in the lateral direction of the lamine. As a result, the "cancellate" laminae overfilled with microfossils turn into thin-laminated structures containing the organic relics only (Plate I, image 4). A similar transition was described in detail by Hofmann (1975), who used Paleoproterozoic stromato-

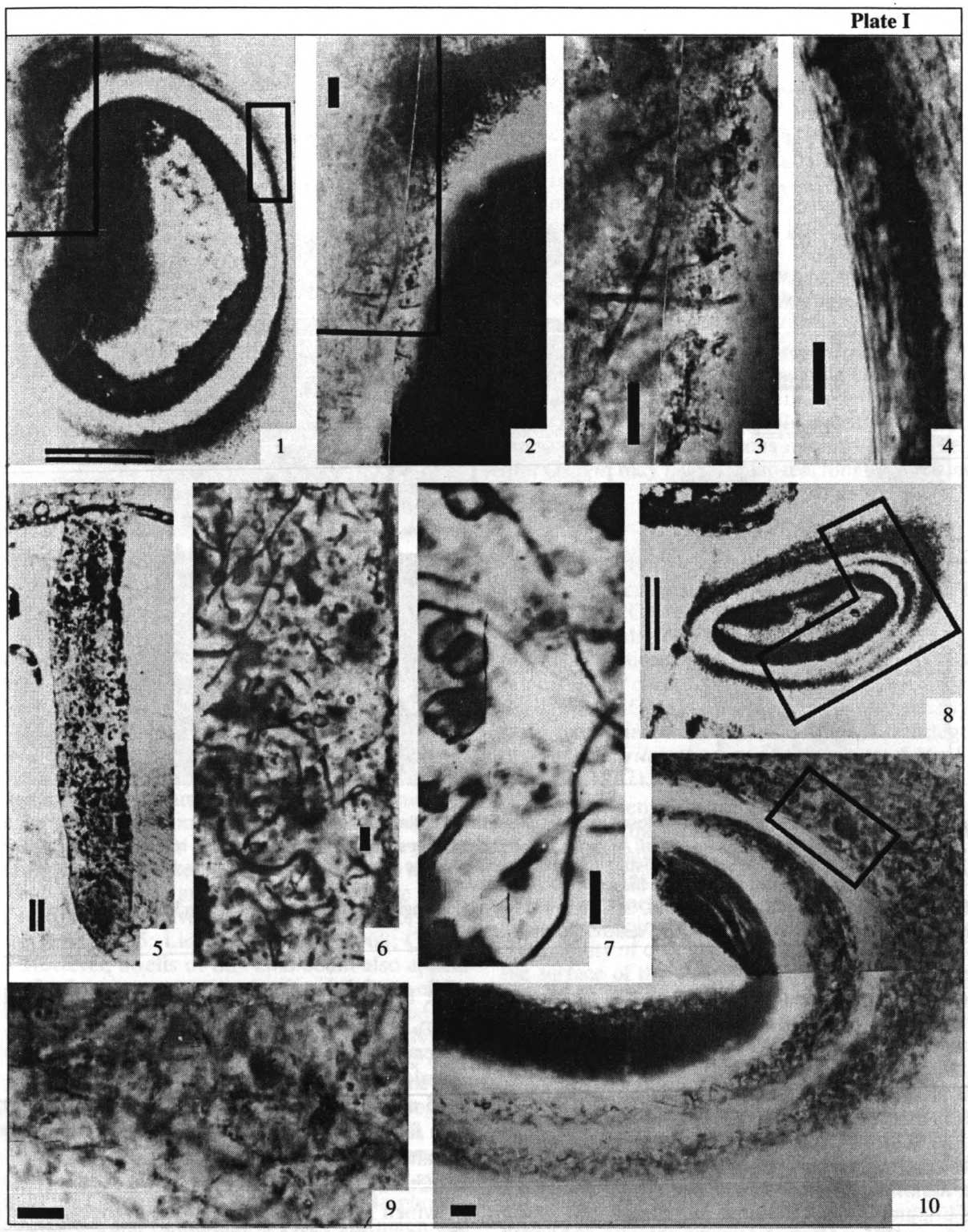


Plate I. Silicified microphytolites and enclosed microfossils of the Gunflint Formation

(1–4) *Osagia* sp. with colonies of filamentous of *Gunflintia minuta*, thin-section 744-96, point 1: (2) is the left rectangular shown in (1); (3) is the rectangular shown in (2); 4 is the right rectangular shown in (1); (5–7) fragment of a stromatolitic lamina with microfossils *Gunflintia* and *Huroniospora*, thin-section 745-96, point 1; (8–10) *Osagia* sp. with laminae composed of the "cancellate tissue"—colonies of *Gunflintia* filaments with spheroids of *Huroniospora*, thin-section 744-96, point 3: (10) is a hexagon shown in (8); (9) is a rectangular shown in 10).

All thin-sections represent Sample 1043-78, Collection 4313 of the Geological Institute of Russian Academy of Sciences. Ordinary scalebar is equal to 10 μm, double bar, to 100 μm.

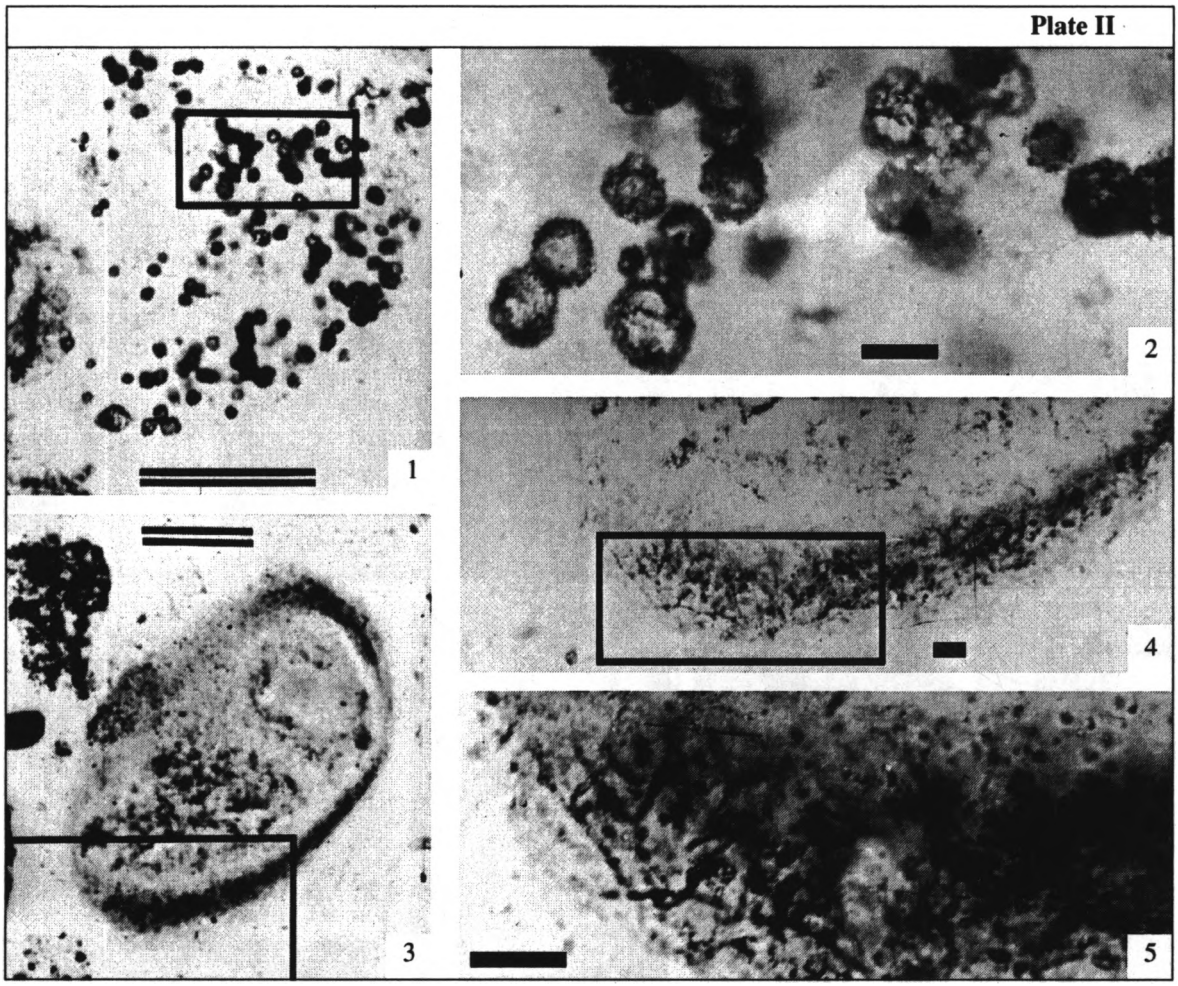


Plate II. Silicified microphytolites and enclosed microfossils of the Gunflint Formation

(1, 2) a colony of *Corymbococcus hodgkissii* Awramik et Barghoorn, thin-section 744-96, point 2: (2) is the rectangular shown in (1); (3–5) *Osagia* sp. with a colony of *Gunflintia* filaments representing the outer lamina of the oncolite, thin-section 743-96, point 1: (4) is the rectangular shown in (3); (5) is the rectangular shown in (4).

All thin-sections represent Sample 1043-78, Collection 4313 of the Geological Institute of Russian Academy of Sciences.

Ordinary scalebar is equal to10 μm, double scalebar, to 100 μm.

lites of northern Canada as an example. In addition to the filaments of *G. minuta* and *G. grandis*, the colonial mats of the dark oncolitic laminae contain sporadic coccoidal unicellular microfossils 1–3 μm in diameter, which are referred to genus *Huronispora* Barghoorn (Plate I, image 9). No microfossils of complex morphology, which were described from the Gunflint microbiota (representatives of genera *Eostrion*, *Kakabekia*, *Xenothrix*, *Archaeorestis*, and others; for review and bibliography see Awramik and Barghoorn, 1977; Schopf and Klein, 1992), have been found in the silicified oncolites studied.

In addition to the oncolites, rare silicified pellets and flakes bearing microfossils are irregularly scattered between columns of the Gunflint stromatolites. Pellets are the rounded or oval nodules 0.2–0.3-mm large, devoid of inner lamination, and variously colored from dark brownish and reddish gray to light gray. Inside them, microfossils form dense mats similar to those observed in the dark oncolitic laminae. Flakes look like angular and poorly rounded microlaminated clasts 0.25 mm thick and up to 1.75 mm long (Plate I, image 5). The fine microstructure of the flakes is determined by the alternation of variously colored thin (0.05–0.1 mm), sometimes not persistent laminae, the darkest varieties of which include remnants of the cyanobacterial mats. The character of lamination suggests these flakes to be fragments of stromatolitic laminae derived from neighboring buildups. Some of these buildups also contain microfossils, but their small dense accumulations were observed only in a single samples of our collection. Some small columnar stromatolites from the same Winston Point section bear much more abundant

microfossils (Awramik, 1976; Awramik and Semikhatov, 1979).

Assemblages of silicified microfossils from pellets and flakes are dominated by mat-forming filaments of *Gunflintia* (Plate I, images 6 and 7), mostly of *G. minuta* and probably rare *G. grandis*, although the latter species is difficult to identify due to the poor preservation of microfossils in these objects. In addition to the filaments of *Gunflintia*, there are spheroids *Huroniospora*, which occur sporadically in pellets and more frequently in flakes, where they are the important components of the enclosed taxonomic association. The pellets also contain frequent star-shaped microfossils *Eoastrion* sp. and single *Eomicrhystridium* sp. spheroids.

In the intercolumnar space of the stromatolites under discussion, the pellets and flakes are associated with loose rounded aggregates of coccoidal microfossils *Corymbococcus hodgkissii* Awramik et Barghoorn (Plate II, images 1 and 2) enclosed in the siliceous microgranular matrix. The aggregates resembling pellets in size and shape represent silicified remnants of colonies of phytoplanktonic microorganisms. They were described many times from the Gunflint microbiota (e.g., Awramik and Barghoorn, 1977; Schopf and Klein, 1992).

Remnants of colonies of the filamentous microfossils, mostly of *G. minuta* and, to a lesser extent, of *G. grandis*, are fragments of the cyanobacterial mats formed mainly by these two taxa. *G.minuta* and *G. grandis* are definitely remnants of hormogonian cyanobacteria, but their affinity to oscillatoriacean or nostocalean microorganisms is a subject of discussion. At the early investigation stage of the Gunflint microbiota, the enlarged cells discovered on the *Gunflintia* trichomes were interpreted by many researchers as akinetes and heterocysts, and the trichomes were compared with representatives of genera *Nostoc* or *Anabaena* (Cloud, 1965; Licari and Cloud, 1968; Cloud, 1976). However, swells of this kind could also appear as a result of the postmortem selective shrinkage of the trichomes of oscillatoriacean *Oscillatoria*- or *Phormidium*-like cyanobacteria lacking them in their lifetime (Golubic and Barghoorn, 1977; Gerasimenko and Krylov, 1983; Krylov *et al.*, 1983). Therefore, the trichomes of *Gunflintia* have an alternative biological assignment even at the order level. Nevertheless, it is evident that these microfossils are remnants of hormogonian cyanobacteria, which formed dense mats at the bottom of the Gunflint basin (Awramik and Barghoorn, 1977; Awramik and Semikhatov, 1979; Hofmann and Schopf, 1983; Lanier, 1989).

The coccoidal microfossils *Huroniospora* found in the *Gunflintia* colonies seem to be remnants of coccoidal mat-dwellers, the microorganisms that inhabited the mats formed by the filamentous forms (Awramik and Semikhatov, 1979). The biological nature of *Huroniospora* spheroids is more disputable than that of the *Gunflintia* filaments. These microfossils are most likely remnants of chroococcalean cyanobacteria (Awramik and Barghoorn, 1977; Golubic and Barghoorn, 1977), the common dwellers in the mats of hormogonian cyanophytes (Golubic, 1976; Knoll and Golubic, 1979; Knoll, 1982). However, the extremely simple morphology of these microfossils does not permit an unambiguous interpretation. They may be remnants of microorganisms of different biological nature, even eukaryotes. The biological interpretation of microfossils *Corymbococcus hodgkissii* Awramik et Barghoorn is more reliable because they are similar in morphology to planktonic cyanobacteria of genus *Aphanocapsa*.

Thus, the microfossils found in the Gunflint oncolites of the group *Osagia* represent the dominant mat-forming hormogonian cyanobacteria and strongly subordinate coccoidal mat-dwellers scattered between filaments. We shall see from below that the established predominance of filamentous microfossils in the Gunflint *Osagia* may have the far-reaching implications.

CONCLUSIONS

When studying catagraphites of the group (formal genus) *Nubecularites* Maslov from the Upper Riphean–Vendian Chichkan Formation of Kazakhstan, Sergeev and Mudrenko (1997) revealed that most of them represent rounded or deformed fragments of colonies of unicellular chroococcalean cyanobacteria *Gloeodiniopsis dilutus* Ogurtsova et Sergeev. This generally confirms the initial Maslov's interpretation of catagraphites *Nubecularites* as rounded fragments of algal films (Maslov, 1937).

The distinctive feature of *Nubecularites* and similar structures is the absence of their synthetic forms. A general similarity between the inner structure of these microphytolites and mats or colonies of coccoidal microorganisms gave rise to a supposition that the absence of their synthetic nodules can be explained by the inability of coccoidal forms to glide and to colonize the surface of the catagraphites unattached to the substrate (i.e., rolling over the bottom). However, the entophysalidacean cyanobacteria are able to escape from accumulating sediments and to recolonize their surfaces by forming abundant mobile gloeocapsoid colonies (Golubic and Hofmann, 1976). This ability allows them, along with filamentous forms, to contribute to the formation of laminated biolites attached to the substrate, i.e., to the stromatolite formation (Golubic, 1976; Knoll, 1982; Schopf and Klein, 1992 and references therein). As to the unattached microphytolites formed in a hydrodynamically active setting, the concentrically laminated and synthetic structures (in particular, *Osagia*) were evidently formed mostly due to the ability of filamentous (hormogonian) cyanobacteria to build dense mats resistible against the environmental hydrodynamics and to twine around the spherical structures encapsulating them entirely or partly. Therefore, the suggestion was made that the origination of concen-

trically laminated buildups of synthetic nodules was controlled by hormogonian (filamentous) cyanobacteria (Sergeev and Mudrenko, 1997), and this is in agreement with the laboratory modeling (Mudrenko *et al.*, 1993).

This suggestion was confirmed by data on the Gunflint microphytolites, which represent concentrically laminated structures of the group *Osagia* forming sometimes the synthetic nodules and containing mat fragments of hormogonian filamentous microfossils of genus *Gunflintia*. Coccoidal forms (spheroids of genus *Huroniospora*), which occur sporadically in these mats, are statistically insignificant remnants of mat-dwellers that did not influence the morphology and microstructure of microphytolites.

Microfossils found in pellets (rounded clasts of silicified carbonate sediments) differ from those in oncolites in both the quantitative and qualitative characteristics. Among the prevailing filamentous forms from pellets, *G. minuta*. evidently plays a leading role. The coccoidal *Huroniosphaera* occur scarcely, but this taxon is associated with abundant benthic complex microfossils of genus *Eoastrion* and with extremely rare planktonic *Eomicrhystridium* forms. By contrast, in oncolites, an assemblage of filamentous forms includes, along with *G. minuta*, the less frequent *G. grandis*, and a greater amount of coccoidal forms, whereas complex forms of microfossils are missing here. This allows us to suggest that assemblages of autochtonous microorganisms participating in the formation of *Osagia* structures were somewhat different from benthic assemblages of carbonate muds.

It is interesting to compare the microfossil compositions in *Osagia* and in morphologically different stromatolites from the deposits of the Winston Point section. First of all, the comparison shows that filamentous *Gunflintia* and coccoidal *Huroniospora* occur in both types of biolites but in different proportions. In oncolites, the abundance of *Huroniospora* is not greater than a few percents of all microfossils found, whereas in stromatolites, its share is 43–46% of statistically significant taxa in the columnar buildups of different morphology and up to 67% in the stratiform buildups (Awramik and Semikhatov, 1979). It is noteworthy that genus *Gunflintia* is represented by a single species (*G. minuta*) in stromatolites, whereas in oncolites this species is associated with rare *G. grandis*. Regretfully, this inference cannot be supported by representative statistic data, because the studied material is limited. Nevertheless, the distinctive compositions of microbiotas that took part in the formation of morphologically different Gunflint stromatolites, on the one hand, and oncolites, on the other, seem to be evident. This was likely a consequence of differences in facial–ecological settings of the corresponding stromatolitic morphotypes and oncolites of the group *Osagia*.

A possible explanation follows from the empirical data on the composition of recent stromatolite-forming assemblages of hormogonian cyanobacteria. In these assemblages, the taxonomic diversity of coccoidal microorganisms (mat-dwellers) is inversely proportional to the harshness of environmental conditions (Golubic, 1976; Knoll, 1982). Therefore, a small amount of coccoidal microfossils *Huroniospora* among the filaments of *Gunflintia* in the studied microphytolites seems to suggest that microphytolites rolling over the basin floor were less favorable for the growth of hormogonian cyanophytic assemblages in comparison with the stable stromatolites attached to the bottom.

The data obtained can be regarded as the first proof of the earlier supposition on the leading role of filamentous cyanobacteria in the formation of Precambrian concentrically laminated microphytolites. In addition, these data reveal for the first time the distinction in compositions of the Gunglint microfossils found in *Osagia* and in morphologically different stromatolites. Although these conclusions can hardly be extrapolated to all Precambrian microphytolites and stromatolites, the presented information is indubitably important for revealing the Proterozoic biolite morphogenesis.

Below follows the description of the Gunflint microphytolites containing silicified microfossils.

DESCRIPTION OF MICROPHYTOLITES

Group *Osagia* Twenhofel, 1919

Osagia spp.

Plate I, images 1–4, 8–10, Plate II, images 3–5

Description. Concentrically laminated nodules usually of rounded, bean-like, and irregularly rounded form have some small surface irregularities; flattened specimens ellipsoidal in sections occur less frequently. The nodule size is 0.3–0.7 mm. The lamination is nonuniform, alternately finer and relatively thicker; it is developed within approximately one third of the nodule radius. The thin lamination is represented by alternating darker (5–8 μm) and lighter (5–15 μm) siliceous laminae variously enriched in residual organic matter and finely dispersed iron compounds. The thicker lamination results from appearance of the relatively thick (up to 30–35 μm) dark-colored "cancellate" laminae composed of dense mats of filamentous microfossils, which contain the rare scattered coccoidal forms (see above). Along these laminae, the microfossils may be worse preserved, less closely packed, and finally disappeared. As a result, the "cancellate" laminae can be replaced by a series of alternating light and dark thin laminae, or by the irregular band of microgranular light gray chert of the same thickness, and this band cannot be differentiated from that composing the thin light-colored laminae. The laminated part of the nodule is seldomly dominated by the "cancellate" laminae. The nonlaminated nucleus of the nodule is usually as big as two thirds of its radius and represented by clasts different in origin, such as cherty sand particle or microphytolite fragments.

Remarks. Among different representatives of the group *Osagia* described in publications, there are several morphologically irregular varieties assigned to different Riphean forms (formal species): *O. pulla* Z. Zhur., *O. libidinosa* Z. Zhur., *O. composita* Z. Zhur., and *O. tolerabila* Jaksh. These forms are much larger than the Gunflint structures (over 1 mm) and devoid of "cancellate" laminae, but they show the well-persistent thicker dark and light laminae. At least some of these distinctive features of the silicified Gunflint and carbonate Riphean forms are consequences of their different preservation mode. Therefore, it is difficult to identify the Gunflint species, although there are no doubts that they belong to the group *Osagia*. We identified them as *Osagia* spp.

Material: several hundreds of specimens of medium preservation degree in 30 thin-sections of the Gunflint cherts; 15 specimens of microphytolites contain the variously preserved microfossils.

ACKNOWLEDGMENTS

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The Lower Ordovician Volkhov and Kunda Horizons and Characteristics of Trilobites and Ostracodes (the Volkhov River, Leningrad Region)

A. Yu. Ivantsov and L. M. Mel'nikova

Paleontological Institute, Russian Academy of Sciences, ul. Profsoyuznaya 123, Moscow, 117647 Russia

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Abstract—Information on lithology, trilobites, and ostracodes from some sections of the Volkhov and Kunda horizons of the Volkhov structural-facies zone (the eastern part of the Leningrad region) is presented. Described in the paper are two types of sections of the Volkhov Formation, which are exposed in a quarry near the Babino Village at the right bank of the Volkhov River, as well as the section occurring in the trench north of the Zvanka Village at the left bank of the Volkhov River, where formations of the Kunda Horizon are exposed in a single succession. The biostratigraphic subdivision on the basis of trilobites (asaphides) is carried out. More than 30 species of ostracode assemblages are revealed. Three new ostracode species are described, and images of most representative ostracode and trilobite species are presented.

Key words: *Volkhov and Kunda horizons, Volkhov structural-facies zone, Leningrad region, lithostratigraphy, trilobites, ostracodes.*

The Ordovician deposits, which are exposed on the surface in the eastern part of the Leningrad region (the southern coast of Ladoga Lake), belong to the Volkhov structural-facies zone (Myagi, 1984a). Type sections of many formations characteristic of the zone are disposed here. Data on the exact location of type sections in the zone, as well as their description, are missing in the literature. The latest studies of the stratigraphy were carried out in the Ladoga Glint area in the 1950s. Lithostratigraphic observations dominate in present-day studies (Dronov, Savitskii *et al.*, 1996), and minor lithologic units (formations and members), vertically and laterally limited, are distinguished in the sections. The correlation of exposures and sections, carried out mostly on the basis of regional erosion surfaces, became extremely exact. This work resulted in the origin of magnificent lithostratigraphic correlation schemes (Orviku, 1960a, 1960b; Popov, Khazanovich *et al.*, 1989). Unfortunately, the stratigraphic importance of the microfauna, collected since the early period of this century, was disregarded, because many samples were referred not to the specific levels of exposures, but to horizons as a whole, or to biostratigraphic zones, the boundaries of which were interpreted differently.

The work presented deals with the lithologic description of some sections of the Volkhov and Kunda horizons, and also with characteristics of trilobites–asaphides (Ivantsov) and ostracode assemblages supplemented with description of a new species (Mel'nikova) from the Volkhov River area (Plates I–IV). The collection of trilobites and ostracodes are stored at the Paleontological Institute, collection nos. 4330 and 4474 respectively.

The scarp of the Ladoga Glint is composed of rocks of the Volkhov, Kunda, and Aseri horizons. Higher layers of carbonate deposits (the Lasnamyagi and Ukhaku horizons) occur in the rear of the glint and are poorly exposed. The sections of the first and second horizons are described in the work.

Boundaries between formations and members are recognized using changes in the rock composition distinguished during the field study. Some boundaries between horizons and beds with trilobites are placed at the regional erosion surfaces, if the latter mark a change in faunal assemblages.

The Volkhov Horizon. A characteristic erosion surface, which separates the Pyaite and Saka members (or submembers, as adopted in the present work) of massive limestone occurring at the base of the calcareous part of the Ordovician sequence, is authorized as the lower boundary of the horizon (*Resheniya ...*, 1987). The upper boundary of the horizon in the Volkhov and Syas' river basins is recognized either at the upper erosion surface of thick-bedded limestone [the lower "white bed" according to Lamanskii (1901, 1905), or the Khamontovo Member] or at the erosion surface at the base of the Sillaoru Formation ("the lower lenticular bed") near the Voibokalo Station and further westward in the glint area. The Volkhov Horizon in the territory under discussion is an equivalent of the Volkhov Formation.

Plate I

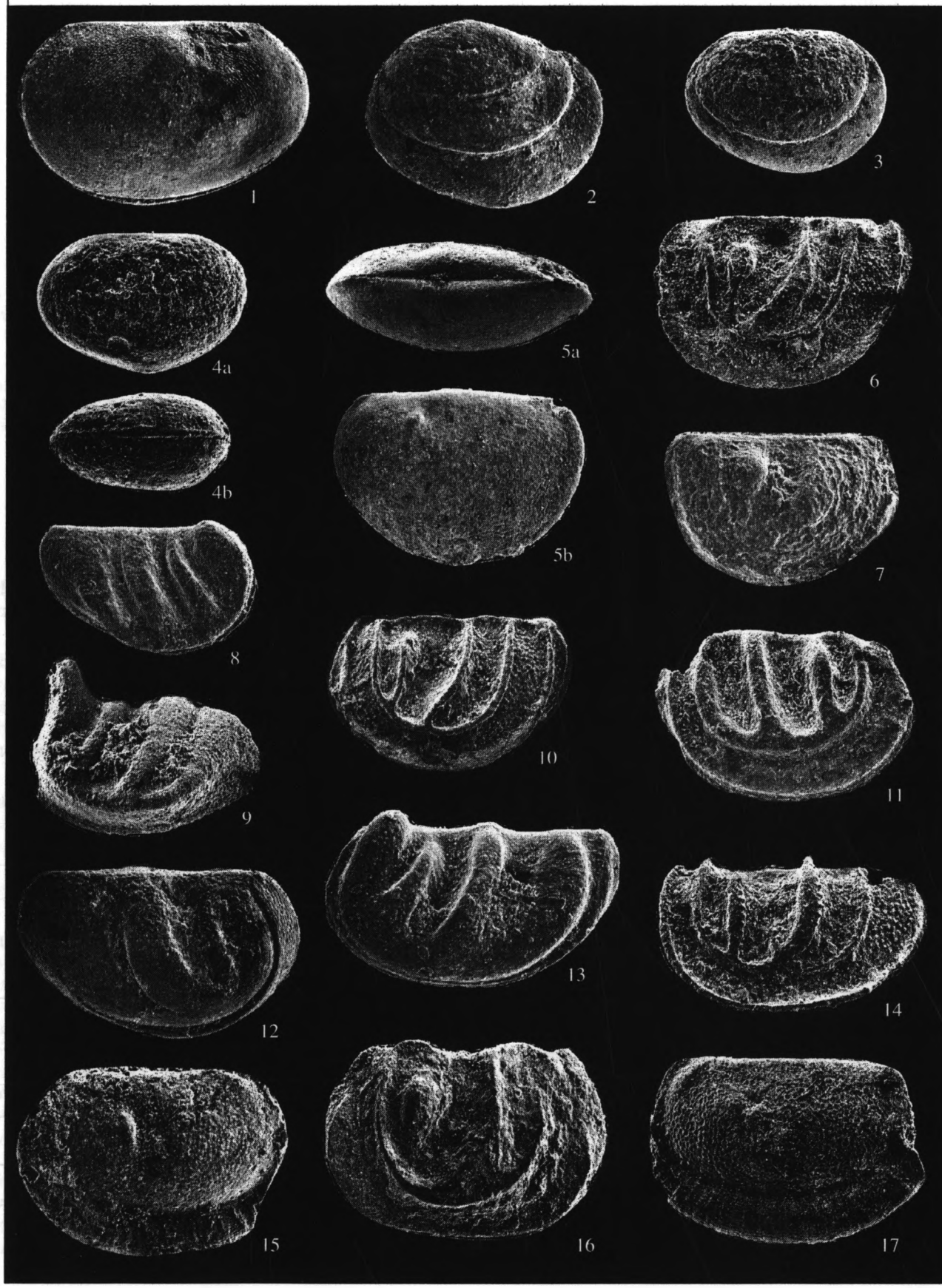


Plate I. Ostracodes from the Volkhov and Kunda horizons

(1–3) *Conchoprimitia gammae* Opik: (1) no. 4474/71, right valve, L = 2.0 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation; (2) no. 4474/72, right valve, L = 1.1 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Frizy" Member; (3) no. 4474/73, right valve, L = 1.65 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Volkhov Horizon, Volkhov Formation, Khamontovo Member. (4) *Miniconchoides minutus* (Hessland), no. 4474/76, shell, L = 0.95 mm: (a) left valve, (b) ventral view; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Ljonna Formation. (5) *Conchoprimitia volkhovensis* sp. nov., holotype no. 4474/79, shell, L = 1.05 mm: (a) dorsal view, (b) left valve; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Zheltyaki" Member. (6) *Rigidella* aff. *plattformis* Schallreuter, no. 4474/82, left valve, L = 0.9 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Zheltyaki" Member. (7) *Rigidella mitis* (Opik), no. 4474/84, left valve, L = 0.45 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Zheltyaki" Member. (8) *Glossomorphites acutus* (Hessland), no. 4474/85, right valve, L = 0.82 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Volkhov Horizon, Volkhov Formation, Khamontovo Member. (9) *Glossomorphites* aff. *grandispinosus* (Hessland), no. 4474/88, left valve from the posteroventral edge, L = 0.7 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Zheltyaki" Member. (10) *Tallinnellina viridis* Schallreuter, no. 4474/89, left valve, L = 0.85 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Zheltyaki" Member. (11, 12) *Brezelina palmata* (Krause): (11) no. 4474/95, female right valve, L = 1.0 mm; (12) no. 4474/99, right valve of the thecnomorph shell, L = 1.01 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Leetse Formation. (13) *Ogmoopsis vesperi* Sarv., no. 4474/102, left valve, L = 0.97 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Zheltyaki" Member. (14) *Piretopsis (Protallinnella) grewingkii* (Bock), no. 4474/104, left valve, L = 1.2 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Frizy" Member. (15) *Laccochilina (Prochilina)* sp., no. 4474/109, left valve, L = 1.52 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Obukhovo Formation, Member 4. (16) *Asteusloffia acuta* (Krause), no. 4474/110, left valve, L = 1.6 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Obukhovo Formation, Member 3. (17) *Laccochilina (Laccochilina)* aff. *kuckersiana* (Bonnema), no. 4474/116, left valve, L = 1.2 mm; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Zheltyaki" Member.

The Volkhov Formation was adopted as a particular lithologic unit by the Interdepartmental Stratigraphic Conference in 1984 (*Resheniya ...*, 1987). The formation is composed of variegated, thick-bedded, lenticular, organic–detrital clayey limestone with frequent clay interlayers and abundant glauconite grains. The Volkhov Formation (within the assumed boundaries) consists of the "Dikari" Member (divided into the Pyaite and Saka submembers), and "Zheltyaki," "Frizy" and Khamontovo members. Units of the glauconite limestone sequences were labelled as "Dikari," "Zheltyaki," and "Frizy" by workers of the Putilovo and Volkhov quarries and introduced into geological terminology by Lamanskii (1905) to become the general informal names of horizons or members of the Volkhov Horizon (Dronov, Savitskii *et al.*, 1993). As the understanding of their lithology did not change since that time, it is reasonable to preserve these original names for members of the Volkhov Formation.

Plate II. Ostracodes from the Volkhov and Kunda horizons

(1) *Hesslandella macroreticulata* (Hessland), no. 4474/119, left valve, L = 0.55 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Volkhov Horizon, Volkhov Formation, Khamontovo Member. (2) *Bolbina major* (Krause), no. 4474/122, shell, L = 0.85 mm: (a) right valve, (b) ventral side; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Ljonna Formation. (3) *Sulcon vadosum* Schallreuter, no. 4474/175, right valve, L = 0.75 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 1. (4) *Longiscula parrectis* Neckaja, no. 4474/128, right valve, L = 0.85 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2. (5) *Collibolbina simplex* (Krause), no. 4474/135, shell, L = 0.7 mm: (a) left valve, (b) ventral side; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Ljonna Formation. (6) *Elliptocyprites ? nonumbonatus* (Hessland), no. 4474/130, right valve, L = 1.1 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2. (7) *Unisulcopleua reticulata* Olempska, no. 4474/132, left valve, L = 0.53 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2. (8) *Eobromidella cicatriosa* (Sarv), no. 4474/138, right valve, L = 1.4 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Obukhovo Formation, Member 1. (9) *Bolbihithes* sp., no. 4474/141, shell, L = 0.5 mm: (a) left valve, (b) ventral side; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Ljonna Formation. (10) *Pinnatulites procerus* (Kummerow), no. 4474/142, left valve, L = 1.75 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village, Kunda Horizon, Obukhovo Formation, Member 4. (11) *Bromidella zvankensis* sp. nov., holotype no. 4474/145, female right valve, L = 1.5 mm: (a) lateral side, (b) interior side of the same valve (dolon is easily seen); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2. (12) *Euprimites (Euprimites) reticulogranulatus* (Hessland), no. 4474/148, right valve, L = 1.5 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 1. (13) *Tallinnella reticulata* Sarv, no. 4474/151, left valve, H = 0.8 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 1. (14) *Tallinnella angustata* (Krause), no. 4474/153, left valve, L = 1.55 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Obukhovo Formation, Member 1. (15) *Easchmidtella ivantsovi* sp. nov., holotype no. 4474/156, shell, L = 1.3 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Obukhovo Formation, Member 4.

Plate II

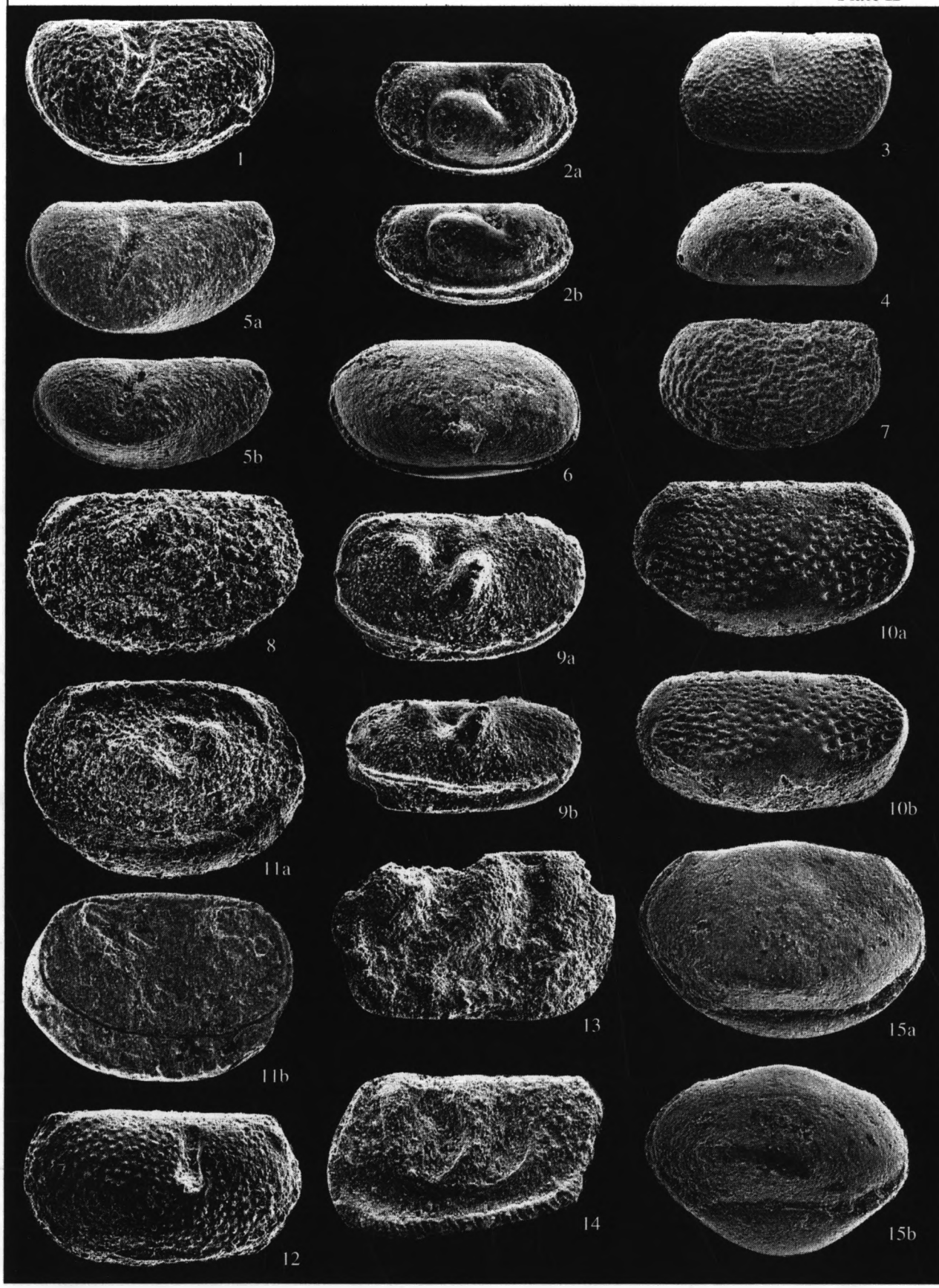


Plate III

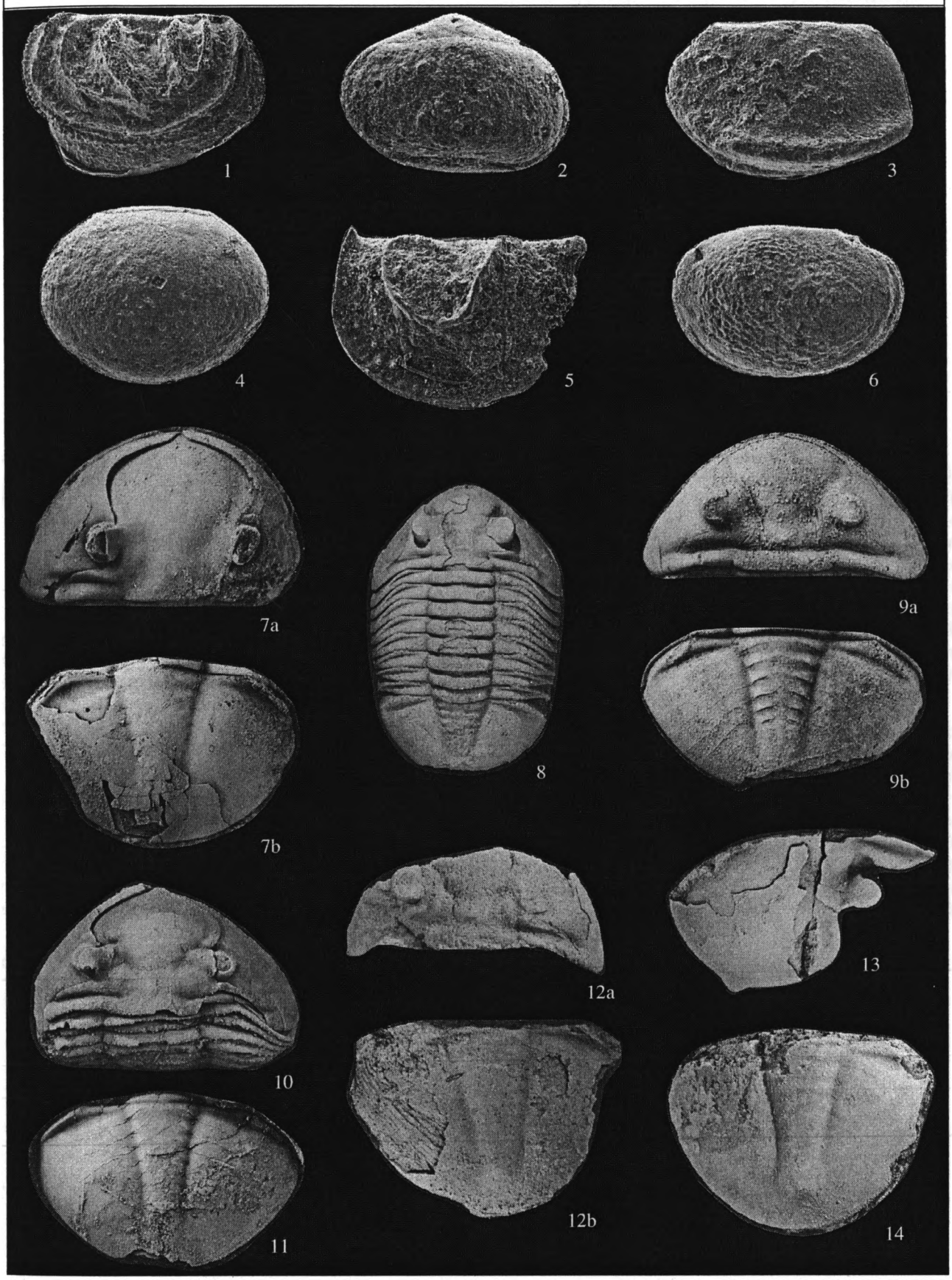


Plate III. Ostracodes and trilobites from the Volkhov and Kunda horizons

(1) *Tallinnellina* sp. A, no. 4474/161, right valve, L = 1.48; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Sillaoru Formation, Member 1. (2) *Aahithis* sp., no. 4474/162, right valve, L = 0.8 mm, the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 1. (3) *Triangulochmidtella* sp., no. 4474/164, left valve, L = 1.05 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 1. (4) *Baltonotella ledaia* Sidaravichiene, no. 4474/166, left valve, L = 1.1 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2. (5) *Lembitsarvella* sp., no. 4474/168, left valve, H = 0.8; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2. (6) *Ahlintella* sp., no. 4474/170, left valve, L = 0.8 mm; the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2. (7) *Asaphus (Asaphus) broggeri* F. Schmidt, no. 4330/9, (× 1.3): (a) cephalon, (b) pygidium; the right bank of the Volkhov River, Babino Village quarry; Volkhov Horizon, Volkhov Formation, "Zheltyaki" Member, 0.65 m above the base. (8) *Asaphus (Asaphus) lepidurus* Nieszkowskii, no. 4330/163, full-size carapace; the left bank of the Volkhov River, northern margin of the Zvanka Village; Volkhov Horizon, Volkhov Formation, "Frizy" Member, 0.65 m below the base of the Kunda Horizon. (9) *Asaphus (Asaphus) expansus* (Wahlenberg), no. 4330/14: (a) cephalon (× 1.2), (b) pygidium (× 1.4); the left bank of the Volkhov River, northern margin of the Zvanka Village; Volkhov Formation, an upper half of the Khamontovo Member. (10, 11) *Asaphus (Asaphus) lamanskii* F. Schmidt: (10) no. 4330/161 (× 1.3), cephalon; the left bank of the Volkhov River, northern margin of the Zvanka Village; Volkhov Horizon, Ljonna Formation, Member 1; (11) no. 4330/13 (× 1.3), pygidium; the left bank of the Volkhov River, northern margin of the Zvanka Village; Volkhov Horizon, Volkhov Formation, an upper half of the Khamontovo Member. (12) *Asaphus (Asaphus) "raniceps"* Dalman, no. 4330/158: (a) cephalon (× 1.3), (b) pygidium (× 1.5); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Sillaoru Formation, Member 2, 0.20 m below the top. (13, 14) *Asaphus (Asaphus) acuminatus* (Boeck): (13) no. 4330/160, cephalon (× 1.2), (14) no. 4330/159, pygidium (× 1.2); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Ljonna Formation, Member 2, 0.10–0.15 m below the top.

The type section of the Volkhov Formation was not identified. The formation, to be more exact the Volkhov Horizon (Balashova and Balashov, 1959), was described on the basis of a series of openings located at the right bank of the Volkhov River near the Izvoz Village and between the Simankovo and Obukhovo villages. The mining of building limestone has long been ceased; the quarries collapsed and overgrew with plants. At present, only the upper part of the formation and its upper boundary are easily accessible for studying in these areas.

A complete section of the Volkhov Formation is stripped of in the stratotype locality by the operating quarry near the Babino Village at the right bank of the Volkhov River. Dmitrovskaya (1991) suggested this section to be a lectostratotype of the formation. The lower boundary of the formation is inadequately distinguished by different researchers—either at the regional erosion surface within the sequence of thick-bedded "Dikari" limestone (*Resheniya* ..., 1987) or along the base of the limestone (Dmitrovskaya, 1989, 1991; Dronov *et al.*, 1993, 1996) similar to the lower boundary of the coeval Toila Formation in Estonia (Myagi, 1984a, 1984b; *Resheniya* ..., 1987; Männil, 1990). We adopt the second viewpoint.

THE VOLKHOV FORMATION IN THE BABINO VILLAGE QUARRY

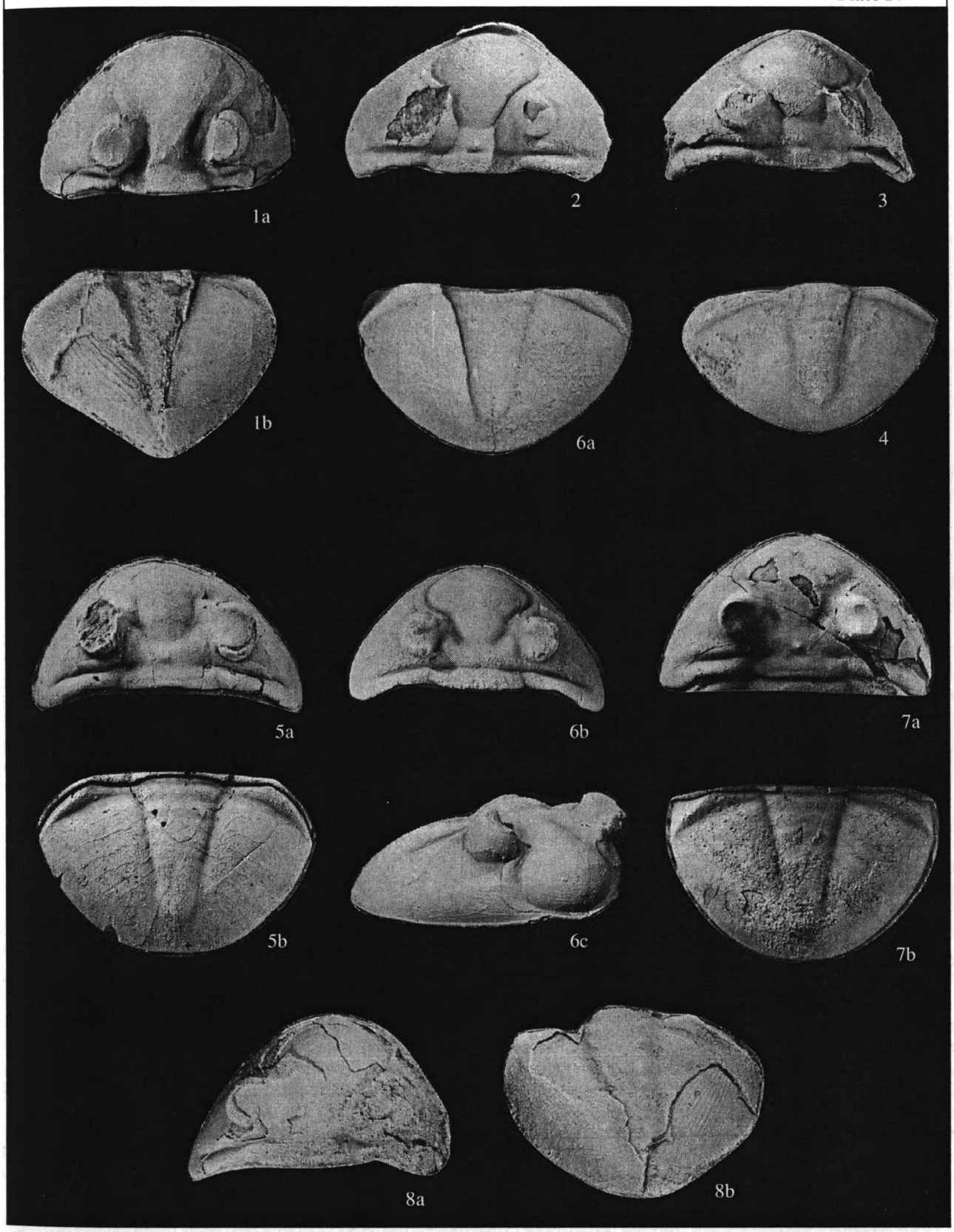
The Volkhov Formation is considered in this work as the limestone beds with glauconite at the base of the carbonate portion of the Ordovician sequence. The lower boundary of the formation is identified at the base of the massive limestone of the "Dikari" Member, whereas the upper boundary is distinguished at the top of the massive stratum of the Khamontovo Member.

Two types of the formation sections, which differ greatly, are distinguished in the Babino Village quarry. The first section typical of the Volkhov time (Fig. 1a) is located in the eastern, northeastern, and western parts of the quarry, where it is composed of limestone and clayey limestone. The second is located in the northern

Plate IV. Trilobites from the Kunda Horizon

(1) *Asaphus (Neoasaphus) minor* F. Schmidt, no. 4330/162: (a) cephalon (× 1.9), (b) pygidium (× 2.3); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, 0.05 m above the base. (2) *Asaphus (Neoasaphus) ingrianus* Jaanusson, no. 4330/167, cephalon (× 1.3); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 1, 0.25 m below the top. (3, 4) *Asaphus (Neoasaphus) laevissimus* F. Schmidt: (3) no. 4330/168, cephalon (× 1.6), the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2, 0.15 m below the top of the Kunda Horizon; (4) no. 4330/166, pygidium (× 1.6); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 2, 0.05 m below the top of the Kunda Horizon. (5) *Asaphus (Neoasaphus) pachyophthalmus* F. Schmidt, no. 4330/35: (a) cephalon (× 1.7), (b) pygidium (× 1.8); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 1. (6) *Asaphus (Neoasaphus) sulevi* Jaanusson, no. 4330/169: (a) pygidium (× 2.3), (b) cephalon (× 2.0), (c) lateral side of cephalon (× 2.2); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Simankovo Formation, Member 1. (7) *Asaphus (Multiasaphus) knyrkoi* F. Schmidt, no. 4330/33 (× 1.2): (a) cephalon, (b) pygidium (x 1.9); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Obukhovo Formation, Member 4, 3.0 m above the base of the "white bed." (8) *Asaphus (Asaphus) striatus striatus* (Boeck), no. 4330/27: (a) cephalon (× 1.3), (b) pygidium (× 1.4); the left bank of the Volkhov River, northern margin of the Zvanka Village; Kunda Horizon, Obukhovo Formation, Member 3, 0.10 m below the top.

Plate IV



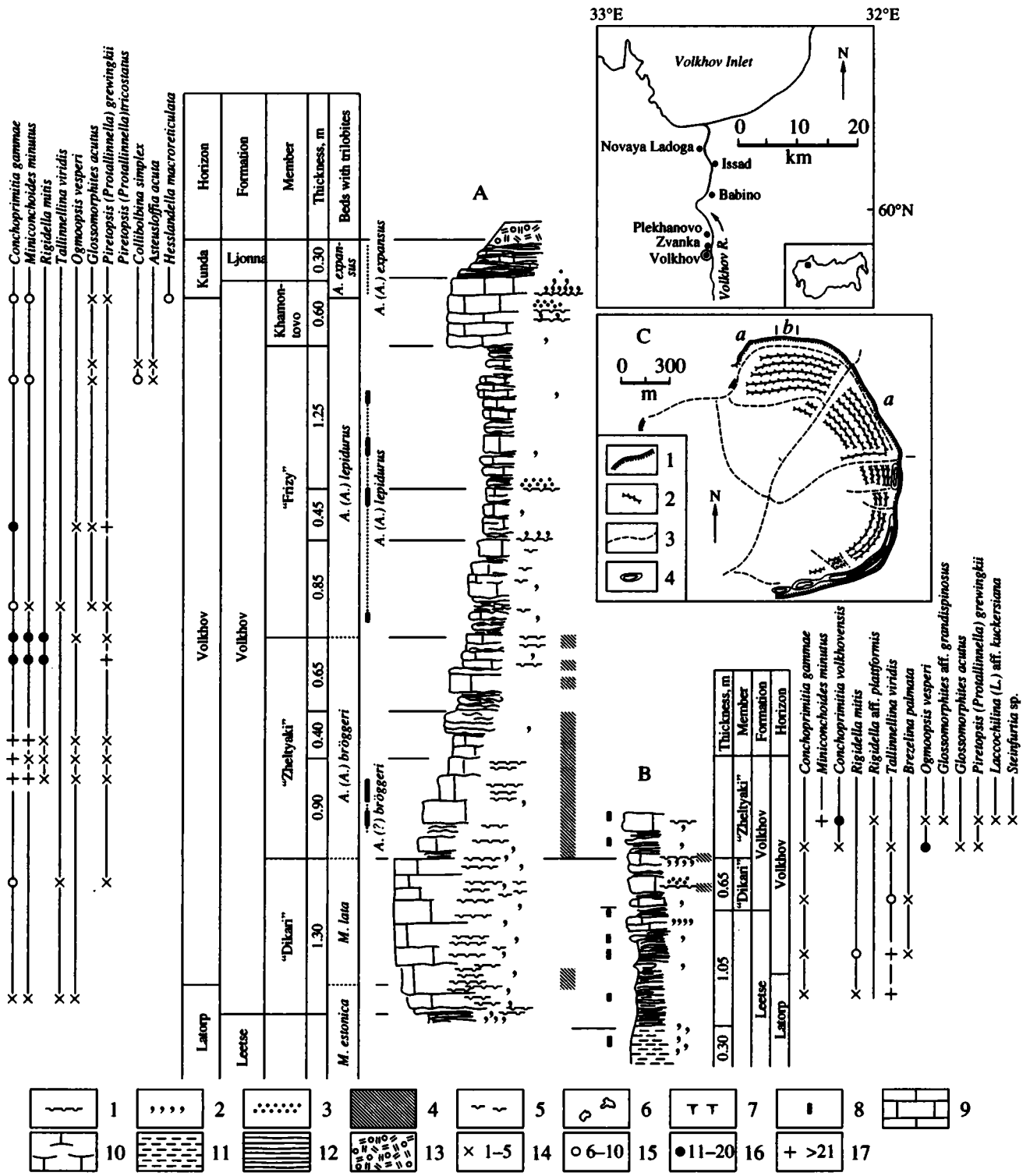


Fig. 1. The lectostratotype section of the Volkhov Formation in the quarry near the Babino Village with indicated occurrences of trilobite and ostracode species: (A) first section; (B) second section; (C) plan of the Babino quarry with indicated (1) active quarry face displaying (a) normal sequence and (b) clayey sequence, (2) waste pile, (3) road, (4) lake. Lithologic features: (1) erosion surface; (2) visible glauconite grains; (3) oolites; (4) red rocks; (5) *Bergaieria*-type tracks filled with glauconite; (6) dissolution cavities; (7) vertical burrows; (8) sampling points; (9) limestone; (10) lumpy limestone; (11) marl; (12) clay; (13) Quaternary deposits; (14–17) number of ostracode specimens.

part of the quarry (Fig. 1b) and consists of clayey limestone and clay. The lower part of the section is represented by calcareous clay and is not comparable with coeval deposits occurring 200 m west and east of it. Judging from blocks of massive micrites common in the spoil bank, one of the so-called "bulges"—small isolated structures of unknown genesis described from the lower Volkhov deposits of the near-glint area—is located here (Dronov and Ivantsov, 1994).

THE FIRST SECTION (A)

Leetse Formation is represented by lilac–gray marl and clay, in places dark green due to abundant glauconite grains. The rocks are exposed in small ditches of water intake pumps and accessible for studying only in the spoil bank. Ostracodes are not found.

Volkhov Formation

"Dikari" Member. Lilac and greenish gray limestone, thick-bedded, medium-grained, organic–detrital, with large and frequent glauconite grains, and frequent erosion surfaces: the lower part of the member 0.6 m thick consists of rusty–brown host rocks coated with coarse-grained detrital glauconite; the middle part is cherry- and gray-colored with a coating of coarse glauconite grains; the upper part 0.25 m thick is most enriched in coarse-grained detrital glauconite. The doubled erosion surface occurs at the level of 0.25 m above the member base. The surface is flat, in places with detached rock fragments 10 cm in diameter and 4 cm high. The underlying rocks are rusty–brown as deep as 10 cm and coated with compact coarse-grained detritus of glauconite. At the surface, there are large burrow-like holes (up to 3 cm in diameter), numerous small attachment structures of echinoderms, and *Bergaueria* burrows. The top of the member represents a typical, even and rusty–brown erosion surface with a scarce detritus of glauconite and abundant burrows of *Bergaueria*. The thickness of the member is 1.3 m.

Ostracodes *Conchoprimitia gammae* Opik, 1935, *Miniconchoides minutus* (Hessland, 1949), *Tallinnellina viridis* Schallreuter, 1993, and *Ogmoopsis vesperi* Sarv, 1959 are found here.

"Zheltyaki" Member. Bed 1. Greenish gray massive fine-grained detrital limestone with frequent brownish yellow and brick-red stains. Medium-sized and large glauconite grains are abundant in a lower part of the bed (0.35 cm); rare glauconite grains are in the upper part (0.55 m). Erosion surfaces are plentiful, one of them occurring 0.20 m of the bed base is very bright rusty brown in color and coated with glauconite detritus. The thickness is 0.90 m.

Bed 2. Brownish yellow limestone fine-grained highly clayey, with clay interlayers and brick–red stains. Medium to small glauconite grains are rare. The thickness is 0.40 m.

Bed 3. Lilac–gray limestone with light green stains and brick-red spots near erosion surfaces; the rock is clayey, fine-grained, and detrital, with clay interlayers. Large and very large (up to 2–3 mm in diameter) glauconite grains are rare, evenly scattered, or located in the form of amorphous accumulations. There are several erosion surfaces with a red, not deep color in the upper part (0.35 m) of the bed. The thickness is 0.65 m. The total thickness of the "Zheltyaki" Member is 1.95 m.

The member yielded all ostracode species typical of the underlying assemblage and associated at this level with *Rigidella mitis* (Opik, 1935) and *Piretopsis (Protallinnella) grewingkii* (Bock, 1867). In addition, the assemblage includes *Conchoprimitia gammae* Opik, 1935 and *Miniconchoides minutus* (Hessland, 1949) representing the background forms.

Trilobites *A. (?) broggeri* F. Schmidt are found in the lower part of the "Zheltyki" Member.

"Frizy" Member. Bed 1. Clayey, lilac and greenish gray, thin- to medium-bedded, fine-grained, detrital limestone with clay interlayers. Medium and large glauconite grains (up to 2 mm in diameter) are abundant, they are usually accumulated in the Bergaueria cavities. The thickness is 0.85 m.

Bed 2. Limestone similar to that of Bed 1, although glauconite grains are here medium in size and abundant. The base and top of the bed correspond to erosion surfaces that do not change the color of underlying rocks and display the *Bergaueria* burrows. Clay occurring above the surfaces is enriched in glauconite (green reference interlayers). The thickness is 0.45 m.

Bed 3. Limestone similar to that of Bed 1; glauconite grains are small (much less than 1 mm in diameter) and rare. Apparent erosion surfaces are absent. The thickness is 1.25 m.

The total thickness of the member is 2.55 m.

The "Frizy" Member comprises the same ostracode species as found in the underlying member.

Khamontovo Member. Light gray, thick-bedded, compact, medium-grained, organic–detrital limestone with several erosion surfaces (more than three, occurring at a distance of 0.20, 0.30, and 0.45 m from the base). Rocks below them are lilac–gray, and small glauconite grains are concentrated above the surfaces, especially above the upper one. The thickness is 0.60 m.

Among ostracodes from the lower part of the member, there are some species which were found before: *Conchoprimitia gammae* Opik, 1935; *Miniconchoides minutus* (Hessland, 1949); and *Piretopsis (Protallinnella) grewingkii* (Bock, 1867). They associate with new species *Piretopsis (Protallinnella) tricostatus* (Sarv, 1963), *Glossomorphites acutus* (Hessland, 1949), *Collibolbina simplex* (Krause, 1891), and *Asteusloffia acuta* (Krause, 1891).

The upper part of the Khamontovo Member contains, above the uppermost erosion surface, the same ostracodes as in the lower part: *Conchoprimitia gam-*

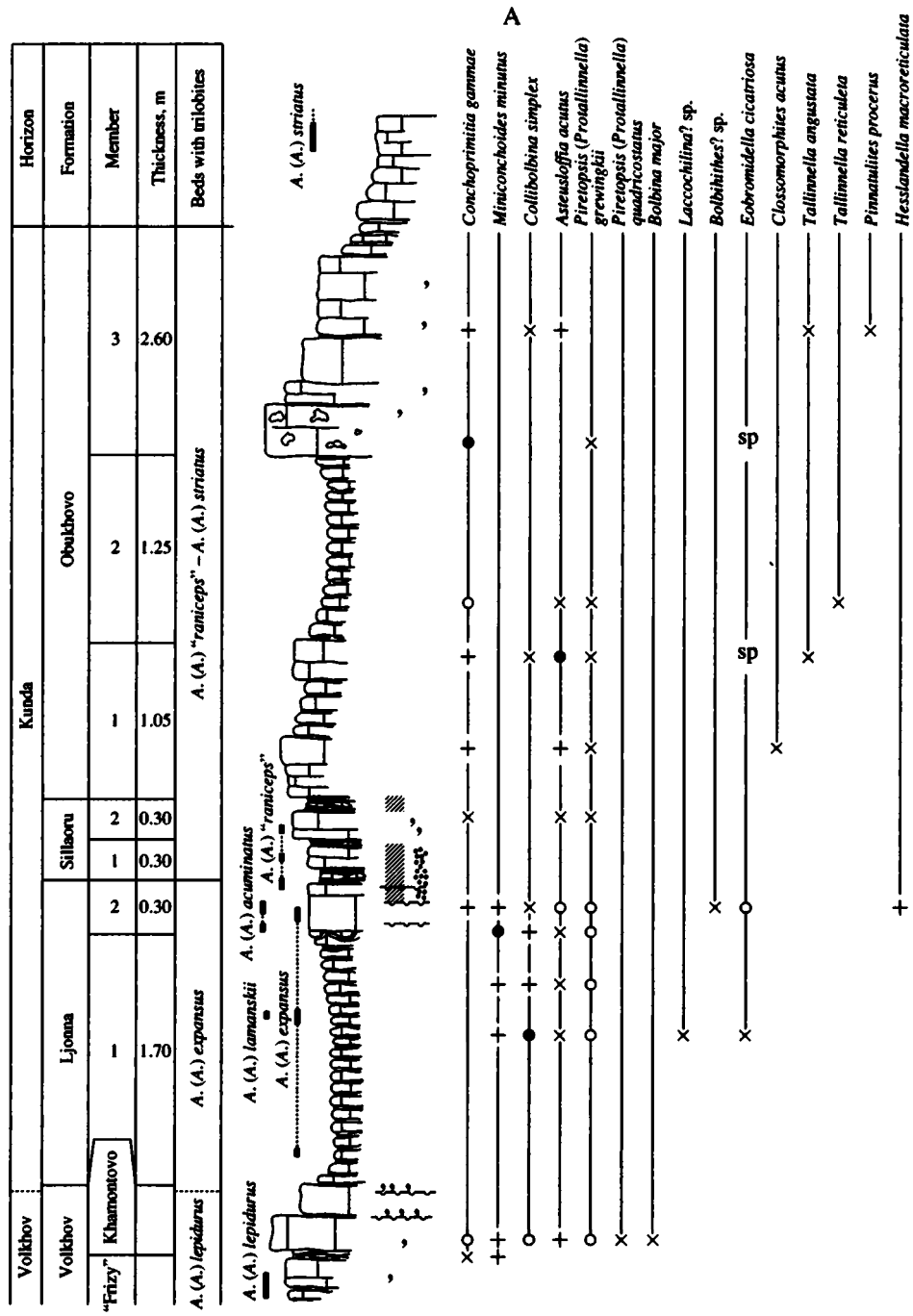


Fig. 2. Section of the Ljonna, Sillaoru, and Obukhovo formations (A) at the left bank of the Volkhov River near the northern margin of Zvanka Village and section of the Obukhovo (continuation), Sinyavino and Simankovo formations (B) at the left bank of the Volkhov River near the northern margin of the Zvanka Village (see Fig. 1 for symbols).

mae Opik, 1935; *Miniconchoides minutes* (Hessland, 1949); *Piretopsis (Protallinnella) grewingkii* (Bock, 1867); *Glossomorphites acutus* (Hessland, 1949); and *Collibolbina* sp. An additional species *Hesslandella macroreticulata* (Hessland, 1949) appears in this interval.

The total thickness of the Volkhov Formation in the first section (A) is 6.40 m.

Deposits of the Volkhov Formation are overlapped by rocks of the Ljonna Formation.

The Ljonna Formation is composed of fine-grained, clayey, lilac, and greenish gray limestone with clay interlayers. Small glauconite grains are abundant at the base of the unit and usually disappear up the section. The visible thickness is 0.30 m.

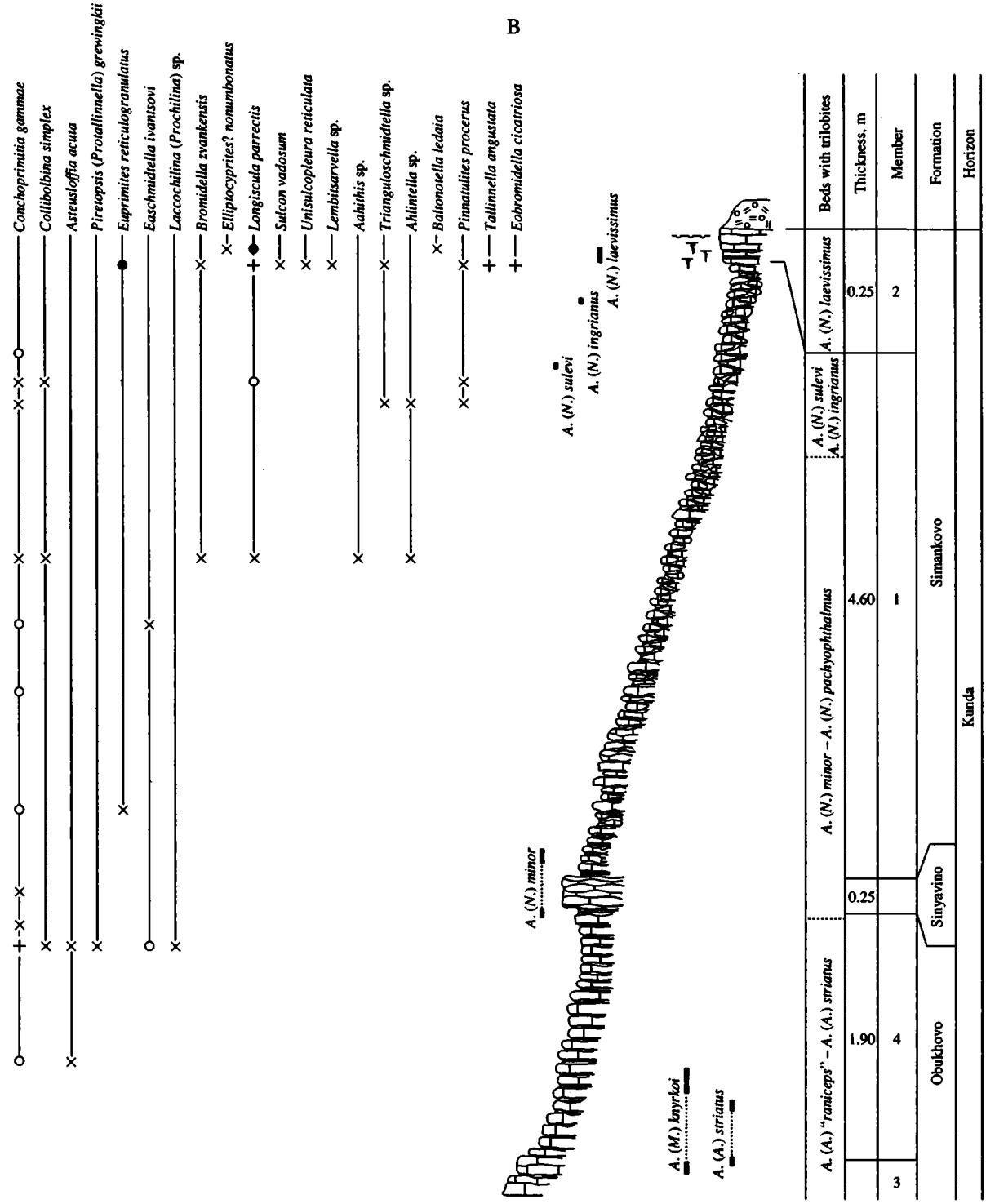


Fig. 2. (Contd).

THE SECOND SECTION (B)

Only the lower part of the section, which differs greatly from the first section (A), is described in this work.

Leetse Formation

Bed 1. Gray to greenish gray marl with glauconite detritus on bedding planes. The visible thickness is 0.30 m.

Bed 2. Lilac–gray clay with calcareous silt-size grains. There is a thin interlayer (0.02 m thick) of gray marl occurring 0.50 m below the bed top. Lenticular interlayers of gray coarse-grained limestone (0.05–0.04 m thick) consisting of echinoderm sclerites bear rare glauconite grains locally emphasizing the cross (?) bedding; the interlayers are observed 0.25 m below the bed top.

A gray marl interlayer 0.10 m thick occurs 0.10 m downward from the bed top; it is overfilled with glauconite grains in the upper part. The thickness of the bed is 1.05 m.

The Leetse Formation yielded ostracodes *Conchoprimitia gammae* Opik, 1935; *Rigidella mitis* (Opik, 1935); *Tallinnellina viridis* Schallreuter, 1993; and *Brezelina palmata* (Krause), 1889.

Volkhov Formation

"Dikari" Member. Alternating lilac and greenish gray, clayey, medium- to fine-grained, organic–detrital, glauconite-bearing limestone and clay beds with lenses composed almost completely of fragments of echinoderm sclerites. There are erosion surfaces occurring at the levels of 0.20 and 0.40 m above the unit base, rusty yellow in color and containing a concentration of thin glauconite detritus. The upper surface correlates with the erosion surface at the top of the "Dikari" Member of the first (A) section corresponding to the "Butok" Bed distinguished by Lamanskii (1905). The thickness of the member is 0.65 m.

Still encountered in the "Dikari" Member are species *Conchoprimitia gammae* Opik, 1935, *Tallinnellina viridis* Schallreuter, 1993 and *Brezelina palmata* (Krause, 1889).

Further up, the rock sequence is like that of the first section.

Identified in the *"Zheltyaki" Member* are ostracodes *Conchoprimitia gammae* Opik, 1935; *Conchoprimitia volkhovensis* sp. nov.; *Miniconchoides minutus* (Hessland, 1949); *Rigidella mitis* (Opik, 1935); *Rigidella* aff. *plattformis* Schallreuter, 1993; *Tallinnellina viridis* Schallreuter, 1993; *Brezelina palmata* (Krause, 1889); *Ogmoopsis vesperi* Sarv, 1959; *Glossomorphites* aff. *grandispinosus* (Hessland, 1949); *Piretopsis (Protallinnella) grewingkii* (Bock, 1867); *Laccochilina (Laccochilina)* aff. *kuckersiana* (Bonnema, 1909); and *Steinfurtia* sp.

BIOSTRATIGRAPHIC TRILOBITE UNITS OF THE BABINO SECTION

The erosion surface corresponding to the boundary between the Latorp and Volkhov horizons is located in the first section 0.25 m above the base of the "Dikari" Member, whereas it is not established in the second section. Index trilobite species of the upper Latorp and lower Volkhov deposits, which are usually represented by *Megistaspis estonica* Tjernvik, *M. lata* (Tornquist) [= *M. (M.) polyphemus* (Brogger)] Nielsen, 1995, have not been found.

Biostratigraphic units of the trilobite fauna in the section near Babino Village are as follows:

Asaphus (Asaphus) broggeri Beds. The species was found 0.30–0.70 m above the top of the "Dikari" Member. Judging from its abundance in the neighboring sections, the species is expected to occur in the lower strata underlying the "Dikari" Member, but no trilobite remains have yet been found in this section inside the "Zheltyaki" Member. Accordingly, the lower boundary of the beds is placed arbitrarily at the "Dikari" Member top.

Asaphus (Asaphus) lepidurus Beds. The first specimens, which likely belong to this species, are encountered 0.15 m above the "Zheltyaki" Member top. Definite remains of this species are found 1.15–2.15 m above the "Zheltyaki" Member top. The lower boundary of the beds probably coincides with the roof of the "Zheltyaki" Member.

Asaphus (Asaphus) expansus Beds. Rare trilobite fragments of this species are found only in debris. Judging from the lithology, they are derived from the Ljonna Formation (lenticular clayey limestone) and from the upper part of the Khamontovo Member (limestone with abundant small glauconite grains).

THE KUNDA HORIZON

The lower boundary of the Kunda Horizon is placed at the upper erosion surface within the Khamontovo Member, above which the mass accumulations of glauconite occur.

The upper boundary of the horizon is recognized by Ivantsov (1993) as corresponding to the erosion surface in the middle part of the Simankovo Formation, as this surface is found to be identical to the doubled erosion surface, which represents the upper boundary of the horizon in the stratotype locality (Orviku, 1960a). Within the Ladoga Glint, the horizon comprises the upper part of the Volkhov Formation (the upper part of the Khamontovo Member) together with the Ljonna, Sillaoru, Obukhovo, Sinyavino and Simankovo (its lower half) formations.

Ljonna Formation. The name was suggested by Myagi (1984a). The type section of the formation is at the left bank of the Ljonna River at a distance of tens of meters from its mouth. We have reduced the range of the formation at the expense of glauconite-bearing limestone, which belongs, in our opinion, to the Volkhov Formation.

Sillaoru Formation. The name was proposed by Myannil and Ryymusoks (1984) to designate the lower sequence of rocks with oolites composed of brown iron compounds (the lower "lenticular bed"). The type section of the formation is at the left bank of the Purtse River near the Sillaoru Settlement in Estonia.

Obukhovo Formation. The formation is distinguished by Myagi (1984a, 1984b) and cited in *Resheniya ...* (1987). The location of the holostratotype is not indicated. Balashova and Balashov (1959) took the right bank of the Volkhov River near the Obukhovo and Simankovo villages as the type region for the Obukhovo Horizon, where it shows the greatest thickness. We suggest the lectostratotype of the Obukhov

Formation, which is exposed at the left bank of the Volkhov River near the northern margin of the Zvanka Village.

Sinyavino Formation. The name "Sinyavino" appears for the first time in *Resheniya* ... (1987) to designate the oolite-bearing member of the Obukhovo Formation. We consider the Sinyavino Member in the rank of the formation, because it represents a direct continuation of the complicated upper sequence of oolitic limestones in the Baltic part of the glint. The eastern, abandoned part of the Putilovo Village quarry is suggested to be the stratotype of the Sinyavino Formation. This part of the quarry best represents the sedimentary sequence adjacent to the Sinyavino Settlement.

Simankovo Formation. The name "Simankovo" was attributed to the upper member of the Obukhovo Formation in *Resheniya* ... (1987). In this work, the member is classified at the rank of the formation of the same stratigraphic range. The spelling is changed (the first letter "o" was replaced by "a") in accord with the name of Simankovo Village. The stratotype of the formation is at the left bank of the Volkhov River, where the ground was excavated during the construction of a new bridge in the Volkhov city.

The formations of the Kunda Horizon are exposed all together at the left bank of the Volkhov River in the trench near the northern margin of the Zvanka Village.

KUNDA HORIZON NEAR THE NORTHERN MARGIN OF ZVANKA VILLAGE

The section begins with the upper part of the Volkhov Formation (Figs. 2a, 2b).

Volkhov Formation

"Frizy" Member: greenish gray wave-laminated limestone with lilac stains, rare clay interlayers, and small glauconite grains. Fragments of *A. (A.) lepidurus* Nieszkowskii are found 0.15–0.25 m below the member top.

Rare Ostracoda species *Conchoprimitia gammae* Opik, 1935 and *Miniconchoides minutus* (Hessland, 1949) are found in the uppermost parts of the "Frizy" Member.

The visible thickness is 0.5 m.

Khamontovo Member (an equivalent to the lower "white bed" of Lamanskii, 1905): massive greenish gray wave-laminated limestone. There are erosion surfaces 0.05 and 0.20 m below the member top, which are marked by accumulations of small glauconite grains above them. The thickness is 0.45 m.

The identified species among Ostracoda fossils are *Conchoprimitia gammae* Opik, 1935; *Miniconchoides minutus* (Hessland, 1949); *Collibolbina simplex* (Krause, 1891); *Asteusloffia acuta* (Krause, 1891); *Piretopsis (Protallinnella) grewingkii* (Bock, 1867); *Piretopsis (Protallinnella) quadricostatus* (Sarv, 1963); *Bolbina major* (Krause, 1892); and *Laccochilina ?* sp.

Ljonna Formation

Member 1: clayey lilac–gray lenticular limestone with abundant clay interlayers; small glauconite grains occur near the member base. *A. (A.) expansus* ssp. 1 is found at the level of 0.25 m above the base, *A. (A.) expansus* ssp. 2 at the level of 1.10–1.20 m above the base, and *A. (A.) lamanskii* F. Schmidt at the level of 1.15 m above the base. The thickness is 1.70 m.

Member 2a: greenish gray wave-laminated limestone. Erosion surfaces with *Bergaueria* burrows occur at a distance of 0.06 and 0.20 m from the member base. Rocks below the lower surface are lilac gray, and those below the upper surface are brownish red. Fragments of *A. (A.) expansus* ssp. 3 are found at the level of 0.10–0.15 m above the base, and remains of *A. (A.) acumunatus* (Boeck) occur at the level of 0.05 m above the base and 0.10 m below the top. The thickness is 0.20 m.

Member 2b: greenish gray clayey limestone with pinkish gray stains around oolite accumulations. Rare accumulations of small oolites of brown iron compounds are observed. An erosion surface with lilac–gray underlying rocks and *Bergaueria* burrows occurs at the member top. The thickness is 0.10 m.

Conchoprimitia gammae Opik, 1935; *Miniconchoides minutus* (Hessland, 1949); *Collibolbina simplex* (Krause, 1891); *Asteusloffia acuta* (Krause, 1891); and *Piretopsis (Protallinnella) grewingkii* (Bock, 1867) are still components of ostracode assemblages of the Ljonna Formation. They coexist here with *Bolbihithes ?* sp., *Eobromidella cicatriosa* (Sarv, 1959), and *Glossomorphites acutus* (Hessland, 1949).

Sillaoru Formation

Member 1: clayey pinkish gray to gray wave-laminated and lenticular limestone with abundant small oolites and interlayers of dark gray clay. Fragments of *A. (A.) "ranicepts"* Dalman are found at a distance of 0–0.05 and 0.20 m above the member base. The thickness is 0.30 m.

Member 2: clayey light gray wave-laminated limestone with rare small glauconite grains. A layer of gray clay with lenses of gray limestone occurs at the member top. The clay is cherry–red around the lenses. The species *A. (A.) "ranicepts"* Dalman is found at the level of 0.10 m above the member base. The thickness is 0.30 m.

Species *Conchoprimitia gammae* Opik, 1935; *Piretopsis (Protallinnella) grewingkii* (Bock, 1867); and *Asteusloffia acuta* (Krause, 1891) are rare in deposits of the Sillaoru Formation, whereas abundant *Hesslandella macroreticulata* (Hessland, 1949) appear in the section for the fist time.

Obukhovo Formation

Member 1: gray limestone with slight lilac-tinted and greenish stains; the rock is wave-laminated, clayey, with very rare small glauconite grains and clay interlayers. The thickness is 1.05 m.

Species *Conchoprimitia gammae* Opik, 1935; *Piretopsis (Protallinnella) grewingkii* (Bock, 1867); *Asteusloffia acuta* (Krause, 1891); *Collibolbina simplex* (Krause, 1891); and *Eobromidella* sp. are still characteristic, and *Tallinnella angustata* (Krause, 1891) and *Tallinnella reticulata* Sarv, 1963 appear for the first time in this unit.

Member 2: gray clayey wave-laminated to lenticular limestone with abundant interlayers of reddish gray clay. The thickness is 1.25 m.

Ostracoda species *Conchoprimitia gammae* Opik, 1935, *Asteusloffia acuta* (Krause, 1891) and *Piretopsis (Protallinnella) grewingkii* (Bock, 1867) are abundant. Individual *Pinnatulites procerus* (Kummerow, 1924) appear.

Member 3: gray wave-laminated limestone with very rare glauconite grains in the lower half of the member. A compact limestone layer, 0.35 m thick, with abundant solution cavities (an equivalent to the upper "white bed" of Lamanskii, 1905) occurs at the member base. The occurrence levels of trilobites are as follows: *Asaphus (Multiasaphus)* sp. 1 1.80 m above the base, *Asaphus (A.) striatus sarsi* Brogger 2.10–2.25 m above the base, and *A. (M.) knyrkoi* F. Schmidt 2.50 m above the base. The thickness is 2.60 m.

Ostracoda species *Conchoprimitia gammae* Opik, 1935; *Piretopsis (Protallinnella) grewingkii* (Bock, 1867); *Asteusloffia acuta* (Krause, 1891); *Collibolbina simplex* (Krause, 1891); and *Tallinnella recticulata* Sarv, 1963 are still typical of the unit, and the additional species *Easchmidtella ivantsovi* sp. nov. appear in it for the first time.

Member 4: gray to pinkish gray clayey wave-laminated to lenticular limestone with abundant interlayers of reddish gray clay. Representatives of the *A. (M.) knyrkoi* F. Schmidt are found at the levels of 0, 0.40, 0.45, 0.50, and 0.60 m above the base; *A. (A.) striatus striatus* (Boeck) occurs in the interval of 0–0.40 m; *A. (N.)* cf. *minor* F. Schmidt is encountered at a distance of 1.95 m from the base, and *A.* (*A.*) sp. nov. at 0.10 m. The thickness is 1.90 m.

Ostracoda species *Conchoprimitia gammae* Opik, 1935; *Collibolbina simplex* (Krause, 1891); *Asteusloffia acuta* (Krause, 1891); *Pinnatulites procerus* (Kummerow, 1924); *Easchmidtella ivantsovi* sp. nov.; and *Laccochilina (Prochilina)* sp. are found in the unit.

Sinyavino Formation: lilac–gray wave-laminated limestone. The thickness is 0.25 m. It should be pointed out that this limestone is arbitrarily referred to the Sinyavino Formation, because oolites of ferric hydroxide characteristic of the formation have not been found here. Accumulations of small oolites of ferric hydroxide, around which the rock turns brown–red, were found only within the upper 0.15 m of the unit section at the left bank of the Volkhov River near the old bridge of the Volkhov city.

Individual representatives of Ostracoda species *Conchoprimitia gammae* Opik, 1935 are found in the formation.

Simankovo Formation

Member 1: lilac–gray clayey lenticular limestone and marl with greenish stains, abundant clay interlayers, and individual pyrite concretions. Trilobites *Asaphus (Neoasaphus) minor* F. Schmidt, *A. (N.) sulevi* Jaanusson, and *A. (N.) ingrianus* Jaanusson occur at the levels of 0.05–0.15, 3.8, and 4.35 m, respectively. The thickness is 4.6 m.

Abundant representatives of Ostracoda found in the member are *Conchoprimitia gammae* Opik, 1935; *Collibolbina simplex* (Krause, 1935); *Euprimites reticulogranulatus* Hessland, 1949; *Easchmidtella ivantsovi* sp. nov., *Bromidella zvankensis* sp. nov., *Longiscula parrectis* Neckaja, 1958; *Sulcon vadosum* Schallreuter, 1993; *Unisulcopleura reticulata* Olempska, 1994; *Lembitsarvella* sp.; *Aahitis* sp.; *Trianguloschmidtella* sp.; *Ahlintella* sp.; *Baltonotella ledaia* Sidaravichiene, 1975; *Pinnatulites procerus* (Kummerow, 1924); *Tallinnella angustata* (Krause, 1891); and *Eobromidella cicatriosa* (Sarv, 1959).

Member 2: compact lilac–gray limestone, light greenish gray in the upper part; the rock is wave-laminated, with slight lumpy jointing and numerous narrow vertical burrows. An indistinct erosion surface with the dark gray color of rocks below and rare traces of burrows is located 0.05 m below the member top. *Asaphus (Neoasaphus) laevissimus* F. Schmidt is found at the levels of 0.05 and 0.15 m above the member base. The apparent thickness is 0.25 m.

Ostracodes *Longiscula parrectis* Neckaja, 1958, and *Baltonotella ledaia* Sidaravichiene, 1975, are encountered in the top portion of the Simankovo Formation, where they coexist with *Elliptocyprites ? nonumbonatus* (Hessland, 1949).

BIOSTRATIGRAPHIC TRILOBITE SUBDIVISION OF THE ZVANKA VILLAGE SECTION

Volkhov Horizon. A. (A.) lepidurus Beds. Only the upper part of the beds is exposed. Remains of *A. (A.) lepidurus* are found at the level of 0.55 m below the upper erosion surface of the Khamontovo Member.

Kunda Horizon. Subdivision of the Kunda Horizon into beds with trilobites, which is suggested in this work, differs substantially from that adopted in *Resheniya* ... (1987). The upper subhorizon corresponding to *A. (Neoasaphus) sulevi* Beds is basically revised. It became a practice to consider three species *A. (N.) sulevi* Jaanusson, *A. (N.) ingrianus* Jaanusson, and *A.*

(M.) knyrkoi F. Schmidt under the common name *A. (N.) sulevi* Jaanusson (= *A. eichwaldi* F. Schmidt). Existence periods of these species were inadequate, and the stratigraphic range of *A. (N.) sulevi* s. 1. is very wide. In addition to stratigraphic ranges of the listed species, it includes the intermediate time intervals between them. The most appreciable time interval is between the stratigraphic ranges of *A. (M.) knyrkoi* F. Schmidt and *A. (N.) sulevi* Jaanusson. The genus *Asaphus* is represented in this interval by the alternating species *A. (N.) minor* F. Schmidt and *A. (N.) pachyopthalmus* F. Schmidt. Moreover, in the upper part of the Kunda Horizon above the range of *A. (N.) ingrianus* Jaanusson, the genus *Asaphus* is represented by the characteristic species *A. (N.) laevissimus* F. Schmidt. Owing to the high abundance and very narrow stratigraphic range of the latter (less than 0.3 m in the most complete eastern sections), the *A. (N.) laevissimus* Beds represent a perfect reference level indicating the top of the Kunda Horizon.

A. (A.) expansus Beds. Balashova and Balashov (1959, p. 133) distinguished the lower boundary of the beds at the level of 3.0 m below the base of the lower lenticular layer in an exposure of the Volkhov River. However, the thickness of the lower Kunda deposits varies and grows along the river from the southern to northern exposures. The first occurrence level of *A. (A.) expansus* is located 1.90 m below the base of the Sillaoru Formation (the lower lenticular layer) in the trilobite–brachiopod coquina accumulated on the upper erosion surface in the Khamontovo Member. This erosion surface should be considered as the lower boundary of the beds and the horizon.

A. (A.) "raniceps"–A. (A.) *striatus* Beds. The first occurrence level of *A. (A.) "raniceps"* is noticed in the oolitic limestone directly above the erosion surface taken for the lower boundary of the Sillaoru Formation. Remains of *A. (A.) striatus* (Boek) (= *A. major* F. Schmidt according to Balashova, 1953) are found only in the third and forth members of the Obukhovo Formation. The lowermost specimens of *A. (M.) knyrkoi* are confined to the upper part of the third member, and this level should be considered as basal for the *A. (N.) sulevi* Beds *s. l.*

A. (N.) minor–A. (N.) pachyophthalmus Beds. The lowermost specimen, which is assumed to belong to the species *A. (N.) minor*, is found 0.05 m below the base of the Sinyavino Formation. The lower boundary of the beds seems to be situated here.

A. (N.) sulevi–A. (N.) ingrianus Beds. The lowermost specimen of *A. (N.) sulevi* Jaanusson is found in the exposure at the level of 0.95 m below the erosion surface of the second member of the Simankovo Formation. However, the studied exposures near the dam of the Volkhov Power Station, where this level is best represented, shows that the lower boundary of the beds should be displaced 0.50 m further below.

A. (N.) laevissimus Beds. Abundant remains of this trilobite species are confined to the middle part of the second member of the Simankovo Formation, to the interval of 0.05–0.15 m below the erosion surface. The correlation with the deposits exposed near the dam of the Volkhov Power Station allows the lower boundary of the beds to be distinguished 0.25 m below this surface.

The Aseri Horizon. A. (N.) platyurus Beds. The upper part of the second member of the Simankovo Formation above the erosion surface (Ivantsov, 1993) should be assigned to the Aseri Horizon, although trilobites have not been found because of a small thickness and area of the exposure.

ANALYSIS OF OSTRACODA DISTRIBUTION THROUGHOUT THE SECTIONS STUDIED

Twenty five samples containing Ostracoda remains have been taken from the Babino quarry and analyzed. The distribution of ostracodes throughout the section is uniform. Stratigraphic analysis (Fig. 1) shows that their taxonomic composition does not vary within the whole Volkhov Horizon, and our attempt to subdivide the Volkhov Horizon on the basis of ostracodes was fruitless. *Conchoprimitia gammae*, *Miniconchoides minutus*, and *Tallinnellina viridis* dominate in the Volkhov ostracode assemblage.

Twenty samples with ostracodes from the section at the left bank of the Volkhov River near the northern margin of the Zvanka Village were also analyzed. The occurrence frequency of ostracodes in this section is variable (Fig. 2). The greatest amount of ostracode remains is confined to deposits of the Ljonna, Sillaoru, and Obukhovo formations, and to the upper half of Member 1 of the Simankovo Formation. The most representative ostracodes for this part of the section are *Conchoprimitia gammae*, *Minoconchoides minutus*, *Asteusslofia acuta*, *Hesslandella macroreticulata*, *Tallinnella angustata*, *Eobromidella cicatriosa*, *Euprimites reticulogranulatus*, and *Longiscula parrectis*. Deposits of the Sinyavino Formation are almost barren of ostracodes. Only individual specimens of *Conchoprimitia gammae* were found here. The detailed subdivision of the Kunda Horizon in the Zvanka Village section is also impossible. Besides, it should be noted that only individual specimens of *Pinnatulites procerus*, the zonal species of the Kunda Horizon (Sidaravichene, 1976), were encountered in the section.

DESCRIPTION OF OSTRACODES

Genus *Conchoprimitia* Opik, 1935

Conchoprimitia volkhovensis Melnikova, sp. nov.

Plate II, Fig. 5

Species name is given after the Volkhov River.

Holotype: no. 4474/79, shell; the Leningrad region, right bank of the Volkhov River, the Babino Village

quarry; the Lower Ordovician, Volkhov Horizon, Volkhov Formation, the "Zheltyaki" Member.

Description. The inequilateral shell is medium in size, postplateous in shape, high, moderately convex. The left valve slightly overlaps the right one along the ventral edge. The hinge line is straight, lies in a mild hollow located at a distance of two-thirds of the dorsal side length closer to the posterior end. The anterior and posterior ends protrude behind the dorsal margin in a similar way. The anterior end is slightly lower than the posterior one. A short, mild groove and an indistinct tubercle in front of it occur on the lateral side of the valves at a distance from the dorsal margin and closer to the anterior end. The middle part of the test is most convex. The posterior end is flattened. The shell surface is smooth.

Size, mm	l	h	w
Holotype no. 4474/79	1.05	0.65	0.4

Variability. The extent of flattening changes slightly along the posterior end.

Comparison. *Conchoprimitia volkhoensis* differs from *Conchoprimitia zonata* (Opik, 1935) in a shorter and milder groove which is located closer to the dorsal margin. The new species differs from other *Conchoprimitia* forms by its dorsal edge line located in a mild groove.

Occurrence: the Lower Ordovician, Volkhov Horizon; the Leningrad region.

Material: about 30 well-preserved shells.

Genus *Easchmidtella* Schallreuter, 1967

Easchmidtella ivantsovi Melnikova, sp. nov.

Plate II, Fig. 15

Species name is given in honor of A.Yu. Ivantsov.

Holotype: PIN RAS, no. 4474/156, shell; the Leningrad region, left bank of the Volkhov River, northern margin of the Zvanka Village; the Middle Ordovician, Kunda Horizon, Obukhovo Formation.

Description. The inequilateral shell is medium in size, ampletous, and highly convex. The left valve overlaps the right one along the free margin; the overlapping is more substantial along the ventral side. The long hinge line lies in a groove formed by elevated median parts of the dorsal margin. Dorsal angles are obtuse, and the anterior angle is much larger than the posterior one. The anterior and posterior ends are even-rounded and equal in height. The lateral side of the both valves steeply bends near the ventral margin to form a wide and flattened ventral side; the bend is more abrupt, carinate on the right valve. The maximum convexity is in the middle part. The shell surface is smooth.

Size, mm	l	h	w
Holotype no. 4474/156	1.3	0.85	0.8

Variability. The extent of expansion of the ventral side is slightly variable.

Comparison. The species differs from the known *Easchmidtella* species by the inflection on the lateral side closer to the ventral margin.

Occurrence: the Middle Ordovician, Kunda Horizon; the Leningrad region.

Material: seven well-preserved shells.

Genus *Bromidella* Harris, 1931

Bromidella zvankensis Melnikova, sp. nov.

Plate II, Fig. 4

Species name is given after the Zvanka Village.

Holotype: no. 4474/145, the female right valve; the Leningrad region, left bank of the Volkhov River, northern margin of the Zvanka Village; the Middle Ordovician, Kunda Horizon, Obukhovo Formation.

Description. Shell is large, ampletous, moderately convex. The dorsal margin is straight and long; the ventral edge is moderately convex. The anterior and posterior ends equally protrude over the dorsal edge line. A wide and deep adductor fossa in the middle of the valve has a wide, rounded median tubercle in front. The fossa is bounded on the rear and from above by the distinct, low, horseshoe-shaped lobe. The dorsal costa is wide; its anterior end bends downward, and the posterior end extends to the posterodorsal angle. The shell surface is ornamented with fine tubercles.

The velar structure of female shells is represented by a wide dolon, which begins below the anterodorsal angle, covers the ventral margin, and extends to the posteroventral edge. Male shells are missing from collection.

Size, mm	l	h
Holotype no. 4474/156	1.5	1.02

Comparison. The new species differs from all *Bromidella* form in a long and convex dolon. Furthermore, it differs from *Bromidella kohtlensis* (Opik, 1937), which has a similar structure of the dorsal rib, in a more pronounced horseshoe-shape lobe.

Occurrence: the Leningrad region; the Middle Ordovician, Kunda Horizon.

Material: two well-preserved valves and several fragments.

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New Data on Cenomanian–Santonian Deposits in the Transaltai Gobi (Mongolia)

G. G. Martinson†, * and N. N. Verzilin**

**Institute of Limnology, Russian Academy of Sciences, ul. Sevast'yanova 9, St. Petersburg, 196199 Russia*
***Institute of the Earth Crust, St. Petersburg State University, Universitetskaya nab. 7/9, St. Petersburg, 199034 Russia*
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Abstract—A section of Cenomanian–Santonian deposits in the Transaltai Gobi (southwestern Mongolia) is described and analyzed. In lithological features and composition of ostracode, mollusk, and charophyte assemblages these deposits are correlative with the Bain-Shire Formation of the Khara-Chutul reference section in the eastern Gobi. The studied counterpart of the Bain-Shire Formation is very complete and can be considered as a reference section for the western Transaltai Gobi.

Key words: *Mongolia, continental deposits, fresh-water mollusks, ostracodes, Cenomanian–Santonian, stratigraphy, correlation.*

In the Mesozoic time, territory of Mongolia represented an intracontinental area of intense accumulation of Jurassic and Cretaceous deposits. The Upper Mesozoic deposits are most widespread in the Gobi region, where they enclose abundant fossil remains of dinosaurs, crocodiles, turtles, fishes, mollusks, ostracodes, conchostracans, and charophytes associated with fossil wood, strobiles, and carpolites of subtropical plants.

Turishchev (1955), Marinov (1957), and Vasil'ev *et al.* (1959) originally subdivided the Upper Cretaceous deposits of Mongolia into the Sain-Shand (Cenomanian) and Bain-Shire (Cenomanian–Turonian) formations. Subsequent geological and paleontological investigations by the Joint Soviet–Mongolian Expedition of the USSR and Mongolian Academies of Sciences resulted in a more comprehensive understanding of basic problems of Mesozoic stratigraphy and correlation in Mongolia. Martinson *et al.* (1969) distinguished the following Upper Cretaceous sedimentary units of Mongolia: Sain-Shand (Albian–Cenomanian), Bain-Shire (Cenomanian–Turonian–Santonian), Barun-Goyot (Campanian), and Nemegetu (Maastrichtian) formations. They believed that the Sain-Shand and Bain-Shire deposits are widespread predominantly in the eastern Gobi, where the Barun-Goyot sediments (the Dzhib-Khalant Formation of this area) are rare. In contrast, the latter and Nemegetu deposits were considered as characteristic of the Transaltai Gobi only. On the basis of these data, one could suppose that sedimentary basins in eastern areas originated earlier than in the Transaltai region, where the process of sedimentation commenced in the Campanian and lasted until the end of the Maastrichtian. The results of Shuvalov (1976), who investigated sediments of the wide Baishin-Tsava basin in the south of the eastern Gobi, were very important and demonstrated that here the Sain-Shand and Bain-Shire formations are overlain by younger sediments bearing fauna and corresponding in age to the Barun-Goyot and Nemegetu formations of the Transaltai Gobi. A little later, Stankevich (1980) described some other outcrops and faunal remains of the Cenomanian–Santonian deposits in the last region. Our study in the Transaltai Gobi resulted in the discovery of a quite complete section of the Bain-Shire Formation rich in fossil ostracodes, mollusks, and charophytes. This section is located at northern slopes of the Mt. Ongon-Ulan-Ula and described for the first time in this article; it can be considered as the reference section for the Transaltai Gobi. Its description is preceded below by a brief review of data on other exposures of the Bain-Shire deposits known in this part of Mongolia.

DISTRIBUTION AREAS OF THE BAIN-SHIRE FORMATION IN THE TRANSALTAI GOBI

When studying the Upper Cretaceous deposits in the Transaltai Gobi, G.G. Martinson, V.F. Shuvalov, and A.V. Sochava encountered some isolated outcrops of variegated sediments dispersed among the dominant Barun-Goyot and Nemegetu deposits and different from them in the lithological features and composition of fossil assemblages. Data on composition of ostracodes (Stankevich, 1980, 1982) and mollusks (Martinson, 1982) from these outcrops showed that their assemblages are characteristic of the Cenomanian–Santonian Bain-Shire Formation.

A section of one of these outcrops located in the southeastern part of the Hermin-Tsav Canyon (Fig. 1)

† Deceased.

was described by Sochava (1975), who attributed it to the Sain-Shand Formation. However, Stankevich considered these deposits as corresponding to the Bain-Shire Formation on the basis of composition of ostracode assemblages recovered from this section consisting of brown clay members intercalated with interbeds of sandstone and variegated clay. The section is crowned with the 16-m-thick member composed of small-pebble conglomerate, gravelstone, and sandstone beds, which are discordantly overlain by the Quaternary conglomerates. According to data of Sochava (1975) and Stankevich (1980), the section in question reveals the following bed succession:

(1) light brown sandstone with carbonate concretions resting on the Paleozoic sandstone (2 m);

(2) variegated clay (3 m);

(3) brown massive sandstone (0.2 m);

(4) brown clay (1.5 m);

(5) greenish to greenish blue clay- and siltstones enclosing carbonate interlayers with abundant deformed casts of *Eucypris hermintsavica* St. (1.5 m);

(6) light blue clay with red spots (0.8 m);

(7) brown nonlaminated clay (2 m);

(8) greenish gray and pink laminated clay with gray carbonate interlayers bearing abundant ostracodes *Cypridea* aff. *profusa* Lub., *Eucypris hermintsavica* St. et Kh., *Lycopterocypris bagataratschensis* St. et Kh., *Cypridea elata* (St. et Kh.), and *Mongolianella* sp. (3.2 m);

(9, 10) brown and yellow clays (14.2 m);

(11) brown clay with ostracodes *Eucypris hermintsavica* St., *Lycopterocypris bagataratschensis* St. et Kh., and *Mongolianella* sp. (8 m);

(12, 13) brown clay with gray and pink sandstone interlayers (13.5 m);

(14) small-pebble conglomerate, gravelstone, and sandstone beds with carbonate concretions (16 m).

Stankevich (1980) also mentioned that a similar 50-m-thick section of variegated clay, sandstone, and sandy clay beds was observed by Shuvalov in the western part of the Hermin-Tsav Canyon. Clay beds from this locality yielded an ostracode assemblage of *Cypridea* aff. *profusa* Lub., *Cypridea* sp., *Eucypris hermintsavica* St., *Mongolianella* sp., *Lycopterocypris bagataratschensis* St. et Kh., *Timiriasevia martinsoni* St., and *Candonella* sp.

Not far away to the northeast, the Bain-Shire deposits were discovered in a small brachyanticline located in the central part of the Shiregin-Gashun depression northward from the Altan-Ula and Nemegetu-Ula mountains. Sochava (1975) erroneously attributed these deposits to the Barun-Goyot and Nemegetu formations. Nevertheless, the diverse fossil fauna from the deposits turned out to be characteristic of the Bain-Shire Formation. This fauna includes ostracodes (Stankevich, 1980) *Mongolianella* sp., *Lycopterocypris*

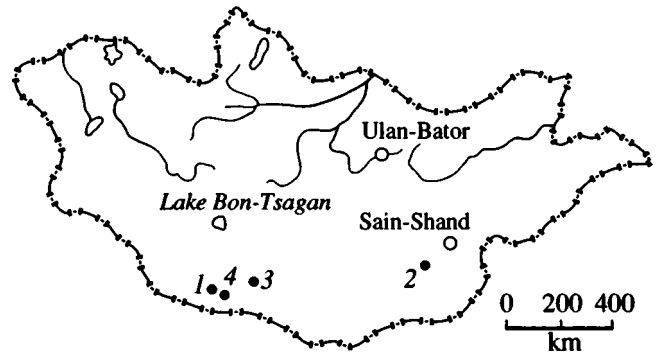

Fig. 1. Geographic scheme of Mongolia with shown localities of the Bain-Shire deposits studied in this work: (*1*) Mt. Ongon-Ulan-Ula; (*2*) Khara-Chutul section; (*3*) Shiregin-Gashun depression; (*4*) section to the southeast of the Khermin-Tsav Canyon.

bagataratschensis (St. et Kh.), and *Cypridea elata* (St. et Kh.) associated with mollusks (Martinson, 1982) *Pseudohyria cardiiformis* (Martins.) and *P. corbiculensis* Martins. Vertebrate remains from these deposits are represented by skeleton fragments of ankylosaur *Syrmosaurus disparoseratus* Maleev and crocodile *Shamosuchus gradilifrons* (Konz.). The deposits also yielded large fossil strobiles of *Araucaria* sp. and carpolites of tropical plants (Krasilov and Martinson, 1982). In the Shiregin-Gashun depression, the Bain-Shire Formation consists of intercalated members of light gray sandstone (with pseudogravel inclusions and lenses at their base) and red clay. Rudaceous deposits are missing in its composition, and the boundary separating these deposits from the Barun-Goyot Formation is unexposed.

In the western Transaltai Gobi, Shuvalov described a section of the northwestern slope of the Mt. Ongon-Ulan-Ula (Fig. 1) that comprises the lower conglomerate member resting on the Paleozoic rocks and the upper member of variegated clays and sandstones. Beds of this section, 40–50 m thick in total, mostly display brown and gray coloration. They yielded large shells of characteristic Bain-Shire mollusks *Plicatotrigonioides chamarinensis* Martins., *P. gobiensis* Martins., *Pseudohyria* cf. *cardiiformis* (Martins.), and *Parreysia* sp. (Martinson, 1982). Ostracodes from the beds (Stankevich, 1980) are represented by *Cypridea* aff. *profusa* Lub., *Mongolianella distilata* (St.), *Eucypris* cf. *hermintsavica* St., and *Cypridea elata* (St. et Kh.). The beds are overlain by the 10-m-thick member of gray–green clay underlying the red deposits of the Barun-Goyot Formation. Unfortunately, this researcher did not publish his own description, and we quote characteristics of the section from the work by Stankevich (1980).

When revising sedimentary formations of the Transaltai Gobi afterward, we encountered a new series of outcrops in the northern part of the Mt. Ongon-Ulan-Ula, the sedimentary sequence of which is much more

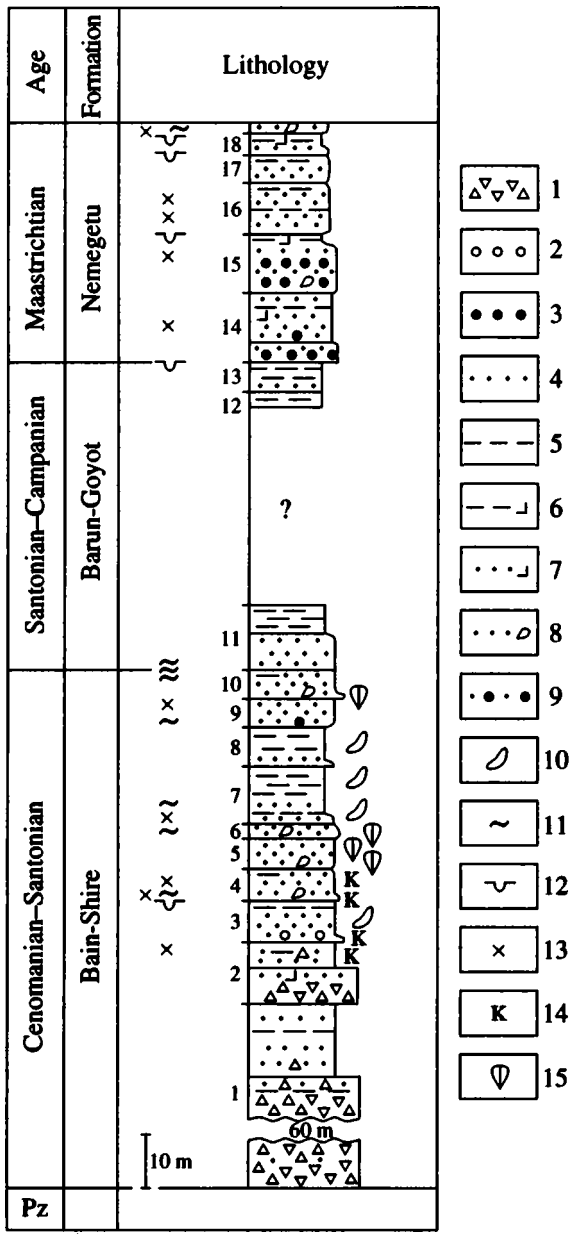

Fig. 2. Section of the Cenomanian–Santonian deposits in the northern slope of the Mt. Ongon–Ulan-Ula: (1) breccia; (2) conglomerate; (3) gravel bed; (4) sandstone and siltstone; (5) clay; (6) clay with carbonate concretions; (7) sandstone with carbonate concretions; (8) pseudogravel or sandstone beds of variable grain size with carbonate and clayey gravely nodules; (9) regularly alternating rocks; (10) remains of ostracodes and charophytes; (11) structural patterns of underwater landslides; (12) neptunian dikes; (13) tracks of crawling animals; (14) bone remains; (15) fossil mollusks (numbers at the left side of the column denote lithological members described in the text).

complete compared to that described by Shuvalov in the northwestern area of the mountain. This sequence, described below, is 162 m thick and can be considered, due to abundant and diverse fossil remains, as the reference section of the Bain-Shire Formation of the Transaltai Gobi.

THE REFERENCE SECTION OF BAIN-SHIRE DEPOSITS IN THE TRANSALTAI GOBI

The section members described below are exposed along the northern slope of the Mt. Ongon-Ulan-Ula (Figs. 1 and 2). Bedding of their deposits is almost horizontal, and basal beds directly rest on the Paleozoic rocks. The difference in altitude between the buried forms of the erosional Paleozoic relief is sometimes greater than 100 m.

(1) The basal gray inhomogeneous breccia, gravel, and sandy gravel beds become colored rusty brown upward and enriched in clay material. The upper third of the member encloses the red clay beds up to 1.5 m thick. At the lower boundary of this unit, we encountered the deformed eggs of dinosaurs. The member is 78 m thick.

In some deep washouts, we observed the onlap relations between Paleozoic rocks and the beds with distinctly evident primary inclinations and pinches. These observations suggest that at the beginning of the Late Cretaceous, the Ongon-Ulan-Ula area represented an erosion zone with intensively dissected relief that supplied the depositional settings with abundant, poorly sorted, and coarse clastic material transported by mud flows.

The yellowish rusty sandstones of the member display frequent gravely inclusions up to 3 cm in size; they enclose interbeds of breccia and red clay. Their thickness is 14 m.

(2) Red inhomogeneous breccias with interbeds of sandstones, nodular carbonate sandstones, and brecciated clay-sandy carbonates; the unit thickness is 7 m.

Light red sandstones with coarse and irregular grain size; they enclose the gravel-size clastics, carbonate nodules, and brecciated lenses or interbeds. Their upper beds are predominantly yellowish pink and marked by tracks of crawling organisms. The top (about 1 m thick) consists of clay incorporating the carbonate nodules and breccias. Remains of dinosaur bones occur in the middle part of the unit 5 m thick.

(3) Dark gray conglomerate with pseudogravel carbonate clastics (rewashed and redeposited carbonate nodules) grades upward in this member into sandstone beds. Fragments of vertebrate bones up to 15 cm in size occur in these rocks. The unit is 0.3 m thick.

Predominantly fine-grained gypsum-bearing sandstones and siltstones with light yellow to pink coloration enclose the red clay interbed with charophytes *Mesochara shuvalovii* Kjan.-Rom. near the top of the unit 13 m thick.

(4) Gray, sometimes pink, medium- to coarse-grained sandstones displaying basal structures of penetration into the underlying clay bed, signs of underwater landslides, and tracks of crawling animals. Locally, they rest on the basal pseudogravel bed consisting of reworked carbonate nodules with fragments of reptiloid bones. The unit is up to 1 m in thickness.

Pink-gray to gray and yellowish brown sandstones, some displaying worm tracks and structural features of underwater landslides; locally they enclose lenses of manganese pseudogravel, where some nodules bear fragments of reptiloid bones up to 10 cm long. The unit is 1.2 m thick.

Yellow and pink medium-grained sandstones; some of them show the cross lamination and deformation structures of underwater landslides. These sediments are 3.5 m thick.

Variegated siltstone grading upward into the red clay; the bed is 0.3 m thick.

(5) Yellow-orange fine-grained sandstone frequently displaying ripple marks and local tracks of crawling animals; the bed is 1.7 m thick.

Yellow to pink sandstones (medium- to fine-grained) intercalated with siltstones; at the base, the rocks locally incorporate pseudogravel lenses with abundant pelecypods *Parreysia gobiensis* Martins., *Sainshandia unegetensis* Martins., *Sainshandia* sp., and *Plicatotrigonioides* sp., and at the level of 1.5 m below their top, there is a small lens of gravely sandstone with clay inclusions and single specimens of gastropods *Lioplax* sp. The unit is 4.5 m thick.

(6) Sandstones of irregular grain size, dark gray at the surface and reddish gray inside; they enclose pseudogravel carbonate inclusions up to 2 cm in diameter and pelecypod assemblage similar in composition to that from the previous bed, but additionally incorporating *Gobiola kharachutulica* Martins. We also observed structural patterns of underwater landslides in this unit 0.2 m thick.

Yellowish pink fine- to medium-grained sandstones with interlayers of siltstone and clay; the unit is 3 m thick.

(7) Bluish light gray medium-grained sandstone with local worm tracks; the bed is 0.6 m thick.

Clay bed brown in its basal portion and variegated above it; the bed yielded charophytes *Mongoliachara paucicostata* Kjan.-Rom. and ostracodes *Candoniella altanica* St., *Cypridea cavernosa* St., and *Eucypris hermintsavica* St. et Kh. The unit is 1.6 m thick.

Bluish gray siltstone with interlayers of mixed silty and sandy material; the bed shows structural patterns of underwater landslides and is 0.8 m thick.

Brown bed of saline clay with gypsum admixture at the base; the recovered charophyte and ostracode species are the same as listed above. The unit is 8 m thick.

(8) Bluish gray sandstone bed, coarse-grained at the base and fine-grained at the top; the thickness of the bed is 1 m.

Crimson-red, slightly saline clay with charophytes *Mongoliachara paucicostata* Kjan.-Rom. and remains of ostracodes *Paracypridea mongolica* Szech. and *Mongolianella* sp.; the intercalated interbeds are composed of bluish light gray fine-grained sandstone. The unit is 6.5 m thick.

(9) Bluish gray coarse-grained sandstone with dispersed gravely grains and structural patterns of underwater landslides; the bed is 0.3 m thick.

Orange-red beds of medium-grained sandstone intercalated with rare interbeds of platy sandstone marked by worm tracks; the rocks are clayey near the top of the unit 5 m thick.

(10) Pseudogravel bed with light red inclusions and large sporadic shells of pelecypod *Plicatotrigonioides gobiensis* Martins. up to 20 cm in size; the bed is 0.5 m thick.

Orange-red to yellow-orange sandstones enclosing rare interbeds of hard platy rocks; the unit is 3.0 m thick.

Orange, medium- to fine-grained sandstone beds intercalated with interbeds of alternating bluish light gray sandstone, siltstone, and red clay, the latter with rare charophyte and ostracode remains; the unit is 2.0 m thick.

The total thickness of the Bain-Shire Formation is 162 m.

Judging from lithological features of the overlying bright orange fine- to medium-grained sandstones and red clays, their sequence 65 m thick corresponds to the Barun-Goyot Formation. These rocks underlie the rocks of the Nemegetu Formation 44 m thick and more diverse in the rock coloration and lithology. The last formation basically consists of beds of light to yellow-gray sandstones of variable grain size alternating with beds of gravel and pseudogravel, and with interlayers of red or variegated clays.

We should note that the boundary between the Bain-Shire and Barun-Goyot formations is conventional to a certain extent, because changes in sediment lithology and in composition of charophyte and ostracode assemblages are very gradual. We believe, however, that the very large bivalves, such as *Plicatotrigonioides*, can be confidently considered as distinctive fossils of the uppermost Bain-Shire horizons, since they are missing in younger continental deposits of Mongolia, China, Thailand, and Laos, but existed in these regions in the Late Cretaceous time (Kobayashi, 1968; Ma Qi-Hong, 1989; and some other publications). Paleobiogeochemical data of Kolesnikov (1982) show that this molluscan fauna characterizes the average annual temperature in the Mongolian Gobi basins as equal to 17–18°C, and that the chloride–carbonate mineralization of waters in the basins was approaching 11–12%.

SPATIAL CHANGES IN COMPOSITION OF BAIN-SHIRE DEPOSITS

The presented data characterize the Bain-Shire Formation in its western distribution area, where it is represented in full measure. Its occurrence directly on the buried erosional forms of the Paleozoic relief was certainly responsible for the greater grain size in the rudaceous basal beds, if we compare them with those over-

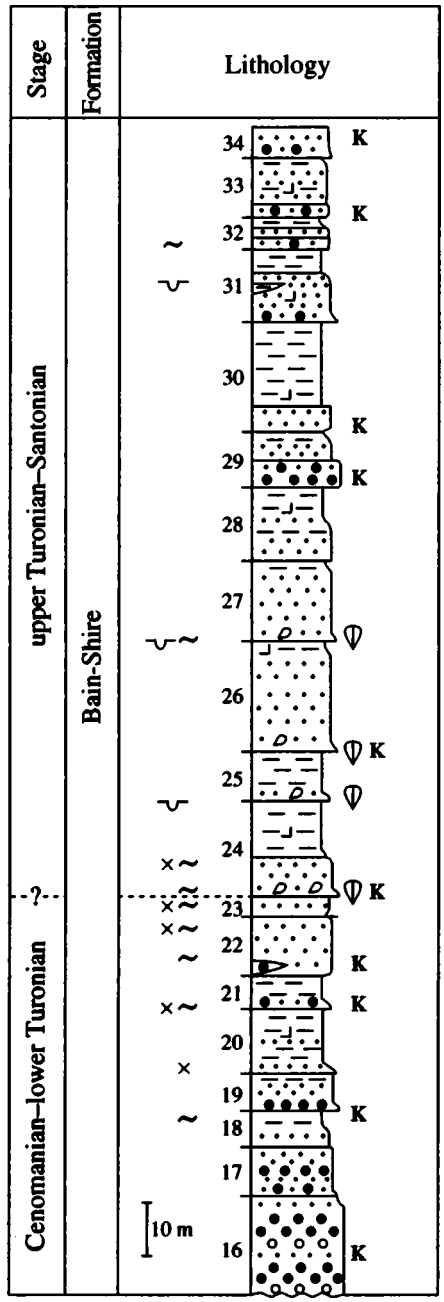

Fig. 3. Section of the Bain-Shire Formation, the Khara-Chutul Ridge (symbols as in Fig. 2; numbers at the left side of the column denote lithological members described by Verzilin and Kalmykova, 1993).

lying the Barun-Bayan Formation in other areas. Nevertheless, the general trend of the grain size increase is characteristic of basal Bain-Shire beds in all regions, where their sequences are congenitally complete. The Khara-Chutul section (Figs. 1 and 3), typical of the eastern distribution areas of Upper Cretaceous deposits in the Gobi region and located 800 km eastward from the section described above, is one of the most interesting in this respect. Its description was published earlier (Verzilin, 1980; Verzilin and Kalmykova, 1993), and we consider here only the most general features of the section.

At the Khara-Chutul locality, the Bain-Shire Formation begins with the bright yellow member of gravelstone, conglomerate, and sandstone beds resting on the erosion surface of the Barun-Bayan Formation. The member is 20 m thick. Deposits of the member yielded remains of Reptilia bones up to 0.5 m long and specimens of fossil wood. Higher in the section, yellow- and light gray sandstone and gravel beds are intercalated with ashy gray to light green clay interlayers, and this unit is overlain by light gray and white sandstone beds 16 m thick in total. Fragmented bones of dinosaurs and turtles are recovered from different levels of this part of the section, and they suggest, according to determinations of L.A. Nesov, the Cenomanian–Turonian age of host deposits. The total thickness of the lower Bain-Shire Subformation is 80 m.

The overlying subformation about 150 m thick consists of alternating members and beds of sandstone and clay. In its lower half, the sandstone members include the pseudogravel beds and lenses in their basal parts, from where the pelecypod shells, bone fragments, and abundant carpolites have been recovered (Krasilov and Martinson, 1982). In the upper part of the same members, the lowermost position is occupied by gravel beds. In contrast to red coloration of clay beds, sandstones are usually light gray in color. The upper boundary of the formation corresponds to an erosion surface separating it from the red beds of the Barun-Goyot Formation. According to Nesov, turtle bones from this part of the Bain-Shire deposits point to its late Turonian–Santonian age. Natural molds of bivalve shells occur at four stratigraphic levels and correspond to typical representatives of the Bain-Shire faunas from many sections. The most characteristic forms among them are *Sainshandia unegetensis* Martins., *S. bainshirensis* Martins., *S. kharachutulica* Martins., *Plicatotrigonioides chamarinensis* Martins., *P. gobiensis* Martins., *Gobiola multicostata* (Barsb.), *G. elliptica* Martins., *Pseudohyria corbiculoides* Martins., *P. cardiiformis* (Martins.), and *Parreysia gobiensis* Martins.

Analyzing the lithological features of the Bain-Shire Formation, we noted the greater abundance of red beds in the Ongon-Ulan-Ula area compared to the Khara-Chutul section, and the fact that basal breccias are missing in the latter. In addition, the Khara-Chutul section is more monotonous in composition, presumably as a consequence of the stable state of provenances which supplied this basin with sedimentary material. At the same time, the distinct transgressive type of the Ongon-Ulan-Ula deposits apparently is evidence of the gradual leveling of the past erosion areas. Nonetheless, basal beds of both localities are coarse-grained, and their deposition was likely related to the interregional changes in intensity of tectonic movements. Despite all these factors, rocks of a single type are usually similar

in their principal lithological parameters and seem to be deposited under quite similar conditions. Some insignificant differences between sections could also be related to a certain climatic contrast between eastern and western areas. For instance, palygorskite is missing from clay minerals of the Khara-Chutul section, and its montmorillonite is of the calc-alkalic type. As opposed to this, the palygorskite admixture is rather frequent in rocks of the Ongon-Ulan-Ula section, and their montmorillonite is represented by alkalic varieties (Verzilin and Kalmykova, 1994). This suggests the lesser mineralization degree of water in the eastern epicontinental basins, which were located in the Bain-Shire period within the area of the present Gobi desert, in contrast to their western counterparts. This could be a consequence of a more arid climate in western Mongolia than in the east of the country.

The very peculiar sections of Bain-Shire deposits are exposed to the southwest and west off the Hermin-Tsav Canyon (Stankevich, 1980). Despite the close proximity of the Paleozoic basement, they are predominantly composed of clayey sediments, and an assemblage of diverse ostracodes recovered from these sediments at the level of 11 m above the Paleozoic rocks is identical to that characteristic to top horizons of the Bain-Shire formation. These data show that the Santonian leveling of the provenance is recorded not only in the Ongon-Ulan-Ula and Khara-Chutul sections, but also in localities between them. In addition, they emphasize the facial variability of the formation and the transgressive type of sedimentation during the respective period.

Paleogeographic maps compiled by Shuvalov (1982, 1994) illustrate the complicated distribution patterns of lakes and lacustrine–alluvial plains of the Santonian time. The Ongon-Ulan-Ula and Khara-Chutul sections described in this work are located, respectively, in the extreme western and eastern parts of this vast area of lacustrine sedimentation. Boundaries of the Bain-Shire lacustrine system were very intricate in shape, the lakes surrounded many islands, and local peculiarities of erosion areas were responsible for the diverse composition of deposits. At the same time, rocks of the studied sections mostly correspond to sediments accumulated at the depth below the basis of wave activity. This is evident from a poor sorting of the terrigenous material, from frequent occurrence of pseudogravel carbonate inclusions in the beds, and from remains of bivalve mollusks buried with conjoint valves.

CONCLUSION

The lithological and paleontological data discussed above show that rocks of the Bain-Shire Formation are rather frequently exposed in the Transaltai Gobi, although in general they are concealed from sight under thick deposits of the Barun-Goyot and Nemegetu formations. In the described section of the northern slope of the Mt. Ongon-Ulan-Ula, the basal Bain-Shire beds are very rudaceous, rest on the erosion relief of the Paleozoic rocks, and correspond to the transgressive trend of sedimentation in the region.

Comparable features of Cenomanian–Santonian deposits in the Transaltai and eastern Gobi regions (the Khara-Chutul section of the latter) are favorable for solving some stratigraphic and correlation problems concerning the Upper Cretaceous sediments of Mongolia.

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Early Paleogene Geological Events in the Western Kamchatka Region

Yu. B. Gladenkov, A. E. Shantser, A. I. Chelebaeva, and V. N. Sinel'nikova

Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Abstract—The new Paleocene stratigraphic scheme of western Kamchatka comprising two, Ulevenei and Chemurnaut groups divided into seven formation and members is discussed. Four horizons marked by marine fossils and four phytohorizons are distinguished. Age and correlation of the defined units are paleontologically substantiated, and, based on these data, a series of paleogeographic maps and a calendar of geological events are compiled. Tectonic, biotic, paleogeographic, volcanic, and climatic events are analyzed.

Key words: *stratigraphy, Paleocene, Kamchatka, paleogeography, volcanism, sedimentation, paleobiota, paleoclimate, correlation.*

INTRODUCTION

Over many years, geologists from the Geological Institute, Russian Academy of Sciences, carry out a complex study of Paleogene and Neogene deposits of the Far East, Kamchatka, Sakhalin, and Koryak Highland, representing the active margin of northeastern Asia. These investigations are aimed at updating stratigraphic schemes and reconstructing the peculiarities of paleogeography and geological development of the region in question. Various sequences were studied in order to elucidate these problems: Eocene and Oligocene sections of western Kamchatka, Paleogene and Miocene sequences in the northeastern areas of the latter (Il'pinskii Peninsula, Korf Bay), Neogene deposits of western (Tigil' and Kavrana areas) and eastern (Karaginskii Island) Kamchatka, and others. Some of these materials were already discussed in a series of publications, including the monograph and summarizing papers (Gladenkov *et al.*, 1985, 1991; Volobueva *et al.*, 1994; Chelebaeva and Shantser, 1985; and others). In these works, much attention was paid, in particular, to substantiation of the regional early–middle Eocene stratigraphic hiatus in the Paleogene sequences, as well as to characteristics of the upper structural complex (the Snatol and Kovachina horizons), and different aspects of sedimentation during the middle–late Eocene were considered. The paper is devoted to data obtained during investigation of reference sections of the lower structural complex of lower Paleogene deposits. These are sections in the Tigil' (Tigil', Snatol, and Kovachina rivers valleys, Utkholok Peninsula coast) and Palana areas (coast from the Cape Kakhtaninskii to the Anadyrka River mouth and Cape Kinkil'skii), where mainly continental facies are developed, and also the Penzhina area (coastal area from the Podkagernaya River Bight on the south to the Chemurnaut Bay and Cape Bol'shoi on the north), where deposits are represented by marine, coastal–marine, and continental facies.

The geographic location of the studied sections is shown in Fig. 1. Based on the materials of study of the lower Paleogene sequences, a manuscript with the analysis and monographic description of marine biota (V.N. Sinel'nikova, A.E. Oleinik, V.N. Ben'yamovskii, S.I. Stupin, N.A. Fregatova) and continental flora (A.I. Chelebaeva, G.B. Chigaeva) is now prepared. In the manuscript, much attention is paid to sedimentation processes (B.V. Polyanskii, K.G. Kazakova) and volcanism (A.E. Shantser, P.I. Fedorov), as well as to tectonics, paleogeography, geological evolution (Yu.B. Gladenkov, A.E. Shantser), and seismostratigraphy (Yu.B. Gladenkov, M.P. Antipov). In this paper, only brief stratigraphic characteristic are given, and the main emphasis is placed on the analysis of early Paleogene geological events in western Kamchatka.

LOWER PALEOGENE STRATIGRAPHY OF WESTERN KAMCHATKA

The section and stratigraphic units. We omit here the history of the development of the stratigraphic schemes of the region, because this issue was discussed in other works (*Geologiya SSSR*, 1964; *Resheniya ...*, 1982; and others). Our investigation of western Kamchatka sections allowed the Ulevenei–Chemurnaut rock complex to be distinguished. The complex includes marine and continental deposits that range in age from the Danian (probably, Maastrichtian) to lowermost Eocene and are overlain by younger members with angular unconformity and hiatus. The complex consists of two units: the lower, Ulevenei and upper, Chemurnaut groups. The former is mainly developed in

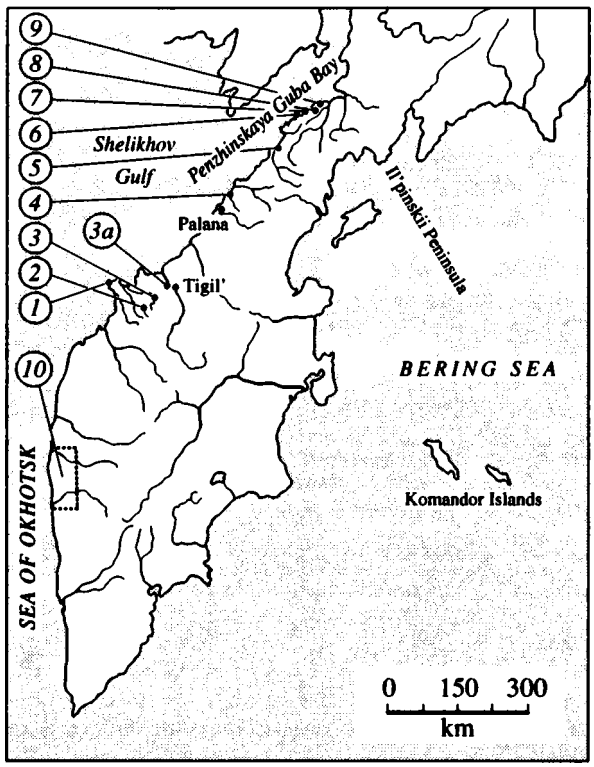


Fig. 1. Location of studied sections: (*1*) Utkholokskii Peninsula; (*2*) Kovachina River; (*3*) Snatol River; (*3a*) Tigil' coal quarry; (*4*) Anadyrka River mouth; (*5*) Shamanka River mouth; (*6*) Tanuinginan Bay; (*7*) area from the Cape Getkilnin to Cape Rebro; (*8*) Evravayam River; (*9*) Chemurnaut Bay; (*10*) area of parametrical boreholes.

the Tigil' area while the latter, in the Penzhina area (Fig. 2).

The oldest layers are discovered in the Tigil' area, where they are united into the Ulevenei Group. Its basal part is represented by the Utkholok volcanogenic unit a few hundred meters thick. It is composed of basalts, andesites, and liparites of a calc-alkaline series and relevant pyroclastic deposits. The upper part of the unit contains lenses and members of tephroids, frequently with plant imprints. The K-Ar age of Utkholok volcanics was determined in the Isotopic laboratory of the State Enterprise "Aerogeologiya" as the late Maastrichtian–early Danian (from 64 ∓ 4 to 60 ∓ 4 Ma). Volcanics are gradually replaced by the terrigenous flora-bearing continental Cape Zubchatyi Formation (700–750 m), which is overlain by the Khulgun Formation (670–675 m). The boundary between them is presumably conformable, albeit their direct contact was not observed. The Khulgun Formation is also terrigenous and includes, along with continental facies with plant remains similar to those from the Cape Zubchatyi Formation, interbeds with brackish water mollusks. The upper part of the Ulevenei Group, which we refer to as the Sosopkhan Formation (about 1000 m), consists of conglomerates, sandstones, siltstones, and argillites, frequently coal-bearing. The rocks bear abundant plant remains, and incorporate several interlayers with marine mollusks and benthic foraminifers.

The continental Anadyrka Formation of the Palana area is likely correlative with the Sosopkhan Formation. All the listed formations are referred to the Danian Stage.

The Chemurnaut Group of the Penzhina area, the lower boundary of which was not observed, is subdivided into three formations. The lower, Getkilnin Formation (to 1000 m) is mostly composed of siltstones with marine mollusks, foraminifers, and rare plant remains. The overlying Kamchik Formation (1000 m) mainly consists of sandstones and siltstones with abundant marine fossils and macrofloral remains.

The upper, Tkapravayam Formation (more than 1000 m) is characterized by a coarser lithology of frequent sandstone and conglomerate members, which are particularly characteristic of the section in the Evravayam River valley where the top layers are exposed (the continental member to 900 m thick). In the Tkapravayam Formation, molluscan and floral remains are particularly abundant, the marine fossils are mainly confined to the lower half of the formation. The Tkapravayam flora is characterized by a high species diversity and a preponderance of large-leaved plants. All the above-discussed formations are referred to the uppermost Danian and, mostly, to Thanetian stages.

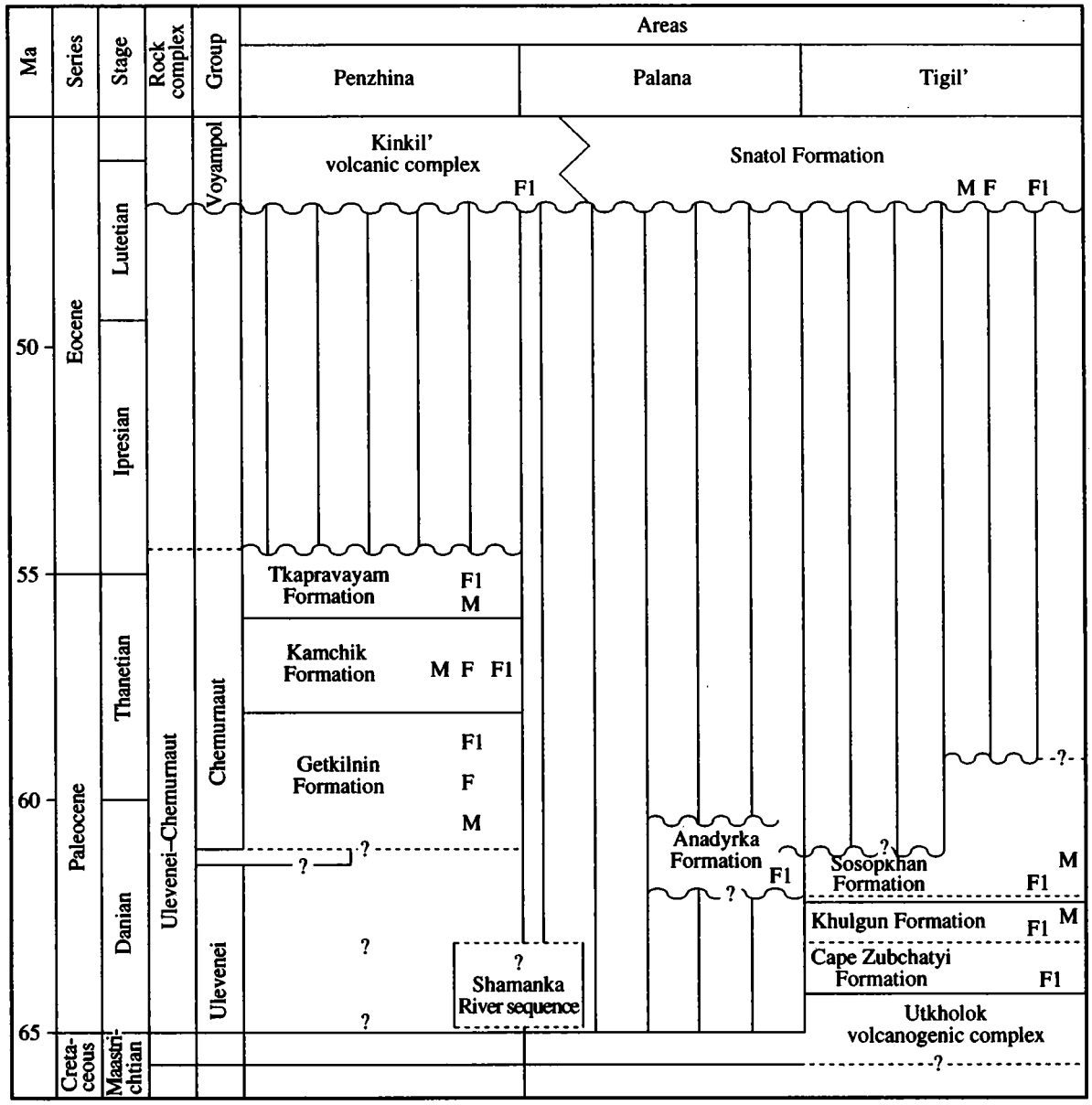
It is not inconceivable that the Cape Kakhtaninskii volcanic complex occurring in the Palana area (the K-Ar age is about 49–50 Ma) can be considered as corresponding to the upper part of the Chemurnaut Group (lowermost Eocene).

The peculiar feature of the lower Paleogene section of Western Kamchatka suggests that it represents, as a whole, a transgressive–regressive sedimentation cycle. The relationships of groups and formations are shown in Figure 2.

Lower Paleogene horizons. The synthesis of stratigraphic data allowed us to revise previous stratigraphic schemes and propose a new subdivision of the sequence into regional stages (horizons). Considering all the available paleontological data, the following stratigraphic succession of horizons can be distinguished.

The lower, *Uluvenei Horizon* corresponds to the Ulevenei Group. Rare marine fossils do not provide grounds for the precise dating, though their Paleocene age is probable (Fig. 3).

By macrofloral remains, the group is subdivided into two phytohorizons: the lower (Utkholok volcanics, Cape Zubchatyi Formation, Khulgun Formation) and upper Ulevenei (Sosopkhan Formation) horizons, whose age spans the Danian and, probably, the earliest Thanetian. As a whole, the temperate mesophilic Ulevenei flora displays a gradual growth in the share in thermophilic elements, and a distinct increase in their role



(M) mollusks, (F) foraminifers, (Fl) flora

Fig. 2. Stratigraphic position of lower Paleogene formations in different areas of western Kamchatka.

at the boundary between lower and upper phytohorizons (at the base of the Sosopkhan Formation). Whereas the share of species with entire leaves in the lower Ulevenei Phytohorizon gradually increases from 8% to 14%, at the base of the upper Uluvenei Phytohorizon it amounts to 20%; higher in the section, it does not exceed 15%. This flora is generally of a more distinct large-leaved type as compared with that from the lower Ulevenei Phytohorizon.

The Getkilnin Horizon is equal in the range to the synonymous formation. It includes three interbeds with mollusks (*Nuculana spheniopsis–Amauropsis, Periploma eodiscus–Eocernina hespericosta–Molopophorus aegunicostatus*, and *Nuculana gabbii–Margarites semiornata* units) and beds with foraminifers (*Silicosigmoilina californica–Kolchidina danica* and a part of *Bulimina rozenkrantzi* zones). According to the marine fauna, the age of the horizon is late Danian–early Thanetian.

Plant remains are rare in these deposits and poorly studied. The Getkilnin Horizon is conditionally correlated with the lowermost Chemurnaut Phytohorizon. It is quite probable that stratigraphically it corresponds to the uppermost Ulevenei Horizon.

The Kamchik Horizon corresponds to the synonymous formation and conformably overlies the Getkilnin

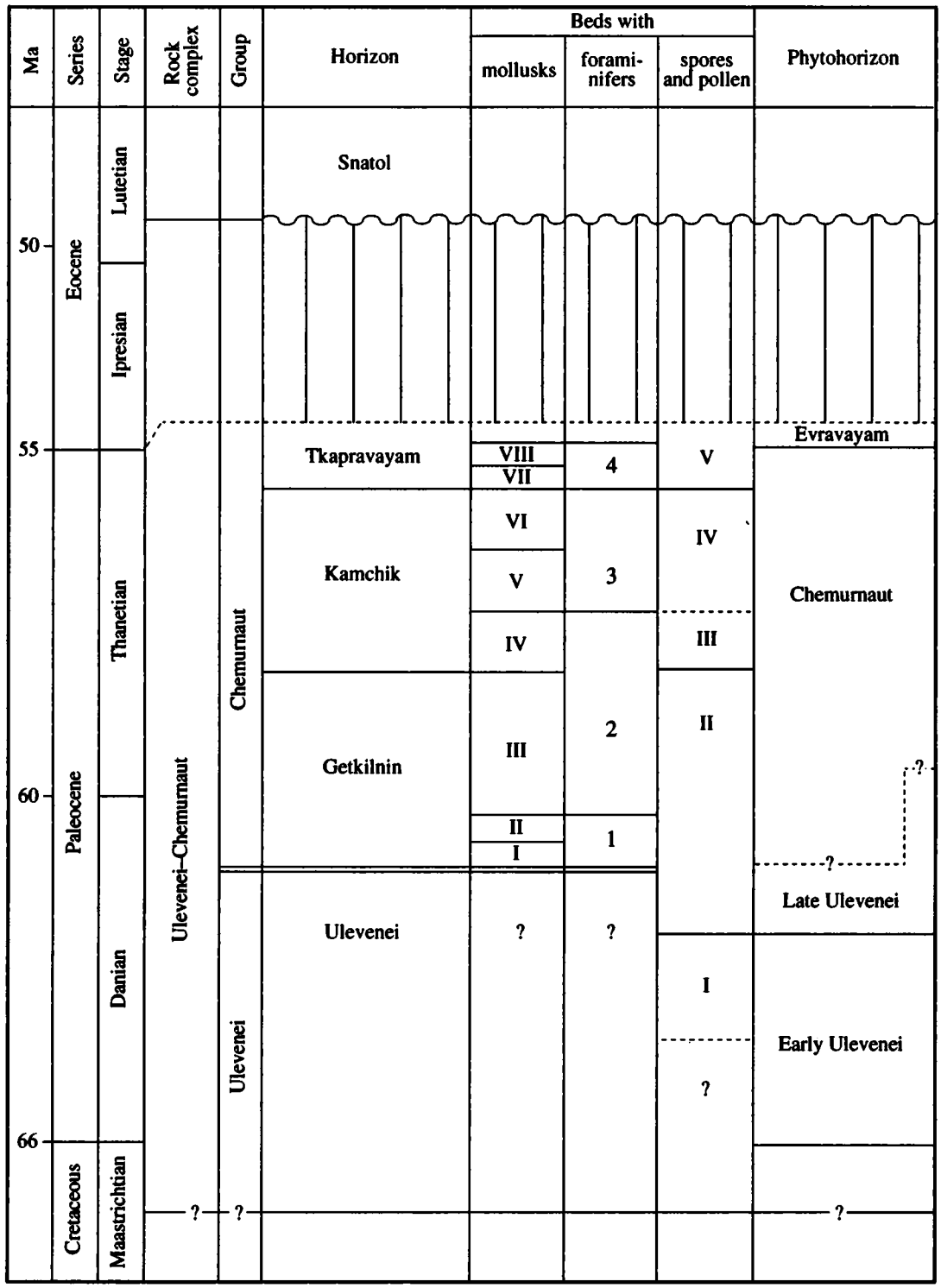


Fig. 3. Lower Paleogene horizons (regional stages) and biostratigraphic units of western Kamchatka.

Horizon. It incorporates three interbeds with mollusks (*Thyasira–Lucina*, *Ostrea ornata–Siphonalia bicarinata*, and *Portlandella kilanskensis–Gilbertina popenoe*) and beds with foraminifers *Bulimina rosenkrantzi* (partly, Getkilnin Horizon) and *Spiroplectammina paleocenica*.

Continental and subcontinental facies are characterized by the large-leaved thermophilic flora, which includes up to 26–27% of dictyoledons with entire leaves. As a whole, floral and faunal assemblages indicate the Thanetian age of the Kamchik Horizon.

The upper *Tkapravayam Horizon* encompasses the synonymous formation, conformably overlies the Kamchik Horizon, and includes two units with mollusks (*Venericardia hornii–Eocernina hannibali* and *Modiolus spesiousus–Potamides penjica* beds) and one bed with foraminifers of *Nonionidea* genus. The flora from the lower part of the horizon is almost similar to that from the middle part of the Chemurnaut Phytohorizon, to which it belongs.

The flora from continental deposits of the uppermost Tkapravayam Formation exposed in the Evravayam River valley forms the autonomous *Evravayam Phytohorizon*. The Evravayam flora displays features indicating decreased atmospheric humidity and increased temperature. The share of entire-leaved species amounts to 36%.

Paleontological assemblages allow the Tkapravayam Horizon to be dated as late (or latest) Thanetian–earliest Ipresian in age. However, the Ipresian flora of Alaska (early Ravenian), corresponding to the terminal phase of the early Paleogene optimum in the North Pacific region, contains up to 65% of species with entire leaves and is more angustifoliate. This allows us to suggest that the Evravayam Phytohorizon flora, the youngest one in the early Paleogene section of western Kamchatka, most likely is Thanetian in age. The distinguished horizons are shown in Figure 3.

Paleoclimatic data may serve as a proxy in stratigraphic correlation. Climatic changes in Kamchatka during the Paleogene depicted from floral remains generally correspond to those established on the basis of planktonic assemblages (Wolfe and Poore, 1982). The early Sosopkhan climatic maximum can be correlated with the Danian optimum at the boundary between P1 and P2 zones, whereas the long-lasting Chemurnaut climatic event most likely corresponds to the Thanetian optimum of the P4 zone.

Data on the marine fossil groups also evidence a sufficiently stable temperature regime during the Ulevenei–Chemurnaut time. The presence of mollusks and foraminifers of the Californian and Tethyan affinity in marine assemblages of western Kamchatka suggests subtropical or alike conditions. Most thermophilic elements occur in the Tkapravayam Horizon. They probably mark the commencement of the peak climatic optimum of the early Paleogene (terminal Paleocene–initial Eocene).

GEOLOGICAL EVENTS

Based on the new Paleogene stratigraphic scheme of western Kamchatka, we analyzed the main aspects of geological history of the region in question and adjacent territories including the following problems: (1) peculiar features of the Cretaceous–Paleogene boundary and related events; (2) distribution of deposition and denudation (the Okhotsk land) areas; (3) volcanic activity; (4) changes in paleogeographic settings and connection of western Kamchatka with oceanic basins; and (5) tectonic position of Kamchatka in the Paleocene. The majority of geological events are tabulated in the form of a peculiar geological calendar—an event-stratigraphy scheme (Fig. 4).

The Cretaceous–Paleogene boundary: peculiar features of tectonics, volcanism, and sedimentation. It is widely accepted that the lower Paleogene almost universally overlies the Cretaceous deposits with a sharp angular unconformity and hiatus, the consequences of the Laramian orogeny in eastern Asia and North America. Most geologists who carried out investigations in western Kamchatka share this viewpoint until present. This concept is grounded on the observed contacts between formations, which display that the continental coal-bearing sequences universally referred to the Paleocene, indeed rest unconformably upon the Upper Cretaceous terrigenous–siliceous deposits, as well as on differently dislocated rocks ascribed to the Upper Cretaceous Irunei Formation and Paleocene sequences.

However, the study of some reference contacts revealed that the Cretaceous rocks are commonly overlain by continental deposits, which lithologically are similar with Paleocene coal-bearing sequences but contain the Eocene paleofloral assemblage.

Recent data also revealed that Paleocene formations proper (the Cape Zubchatyi, Khulgun, and Sosopkhan formations) form the early Paleocene transgressive–regressive sedimentation cycles, whereas similar Eocene deposits usually compose the basal part of the following Eocene–Oligocene transgressive cycle.

We observed the Upper Cretaceous–Paleocene contact in the sole locality, in the Anadyrka River mouth (Palana area). There, the volcanogenic–sedimentary marine Napana Formation (dated by rare radiolarians as the Campanian–Maastrichtian in age) is overlain, with a hiatus but without apparent angular unconformity, by the continental Anadyrka Formation whose age is determined by flora as late Danian–Thanetian. At the same time, some structures in western Kamchatka show a significant formational similarity of terrigenous facies of both the Cretaceous and Paleogene ages. For instance, based on the inoceram find, the Paleocene age of the "Napana" deposits in the Moroshechnaya River basin (western part of the Tigil' uplift) was revised in the course of recent mapping, and they are now dated as Late Cretaceous in age. Inocerams were also found in the distribution area of the lithologically uniform, sufficiently silty "Getkilnin" sequence in the Pylgavayam and Pamanka river basins (western side of the Lesnovskii uplift); in northern areas, lithologically similar rocks are dated back to the Paleocene (stratotype of the Getkilnin Formation).

It should be noted that the subarkose (mainly quartz–feldspar) composition is characteristic of both the lower Paleocene (Cape Zubchatyi Formation, Getkilnin Formation in the stratotype) and Upper Cretaceous sequences, e.g., of the Lesnovskii Formation (in

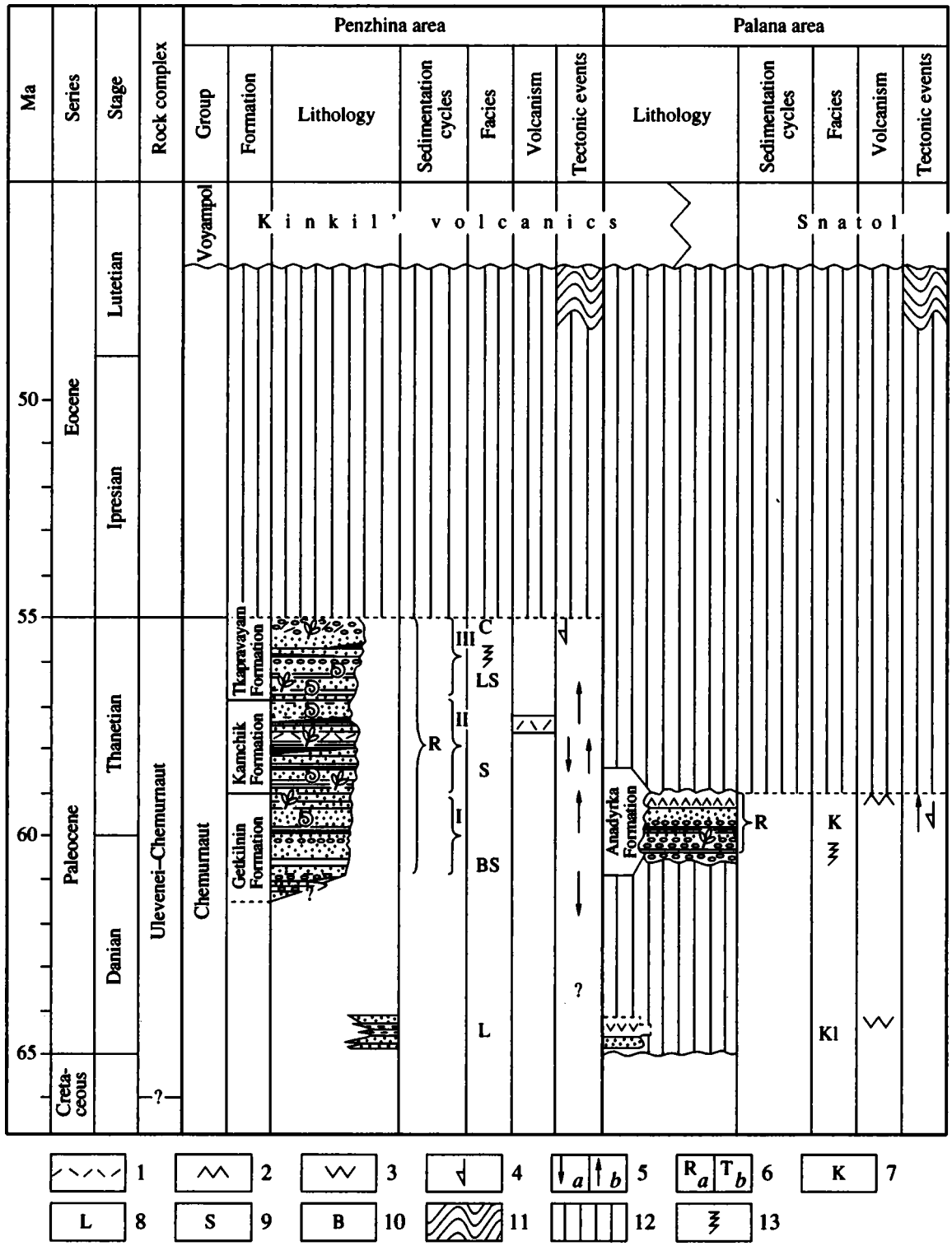


Fig. 4. Scheme of the Paleocene event stratigraphy for western Kamchatka: (1) volcanic ash interlayers; (2) mainly acid volcanism; (3) mainly basic volcanism; (4) graben formation against the background of uplifting; (5) vertical movements: (*a*) subsidence, (*b*) uplifting; (6) sedimentation cycles: (*a*) regressive megacycles, (*b*) transgressive megacycles; (7–11) macrocycles comprising (7) continental deposits, (8) littoral deposits, (9) sublittoral deposits, (10) bathyal deposits, (11) folding; (12) hiatuses; (13) avalanche sedimentation.

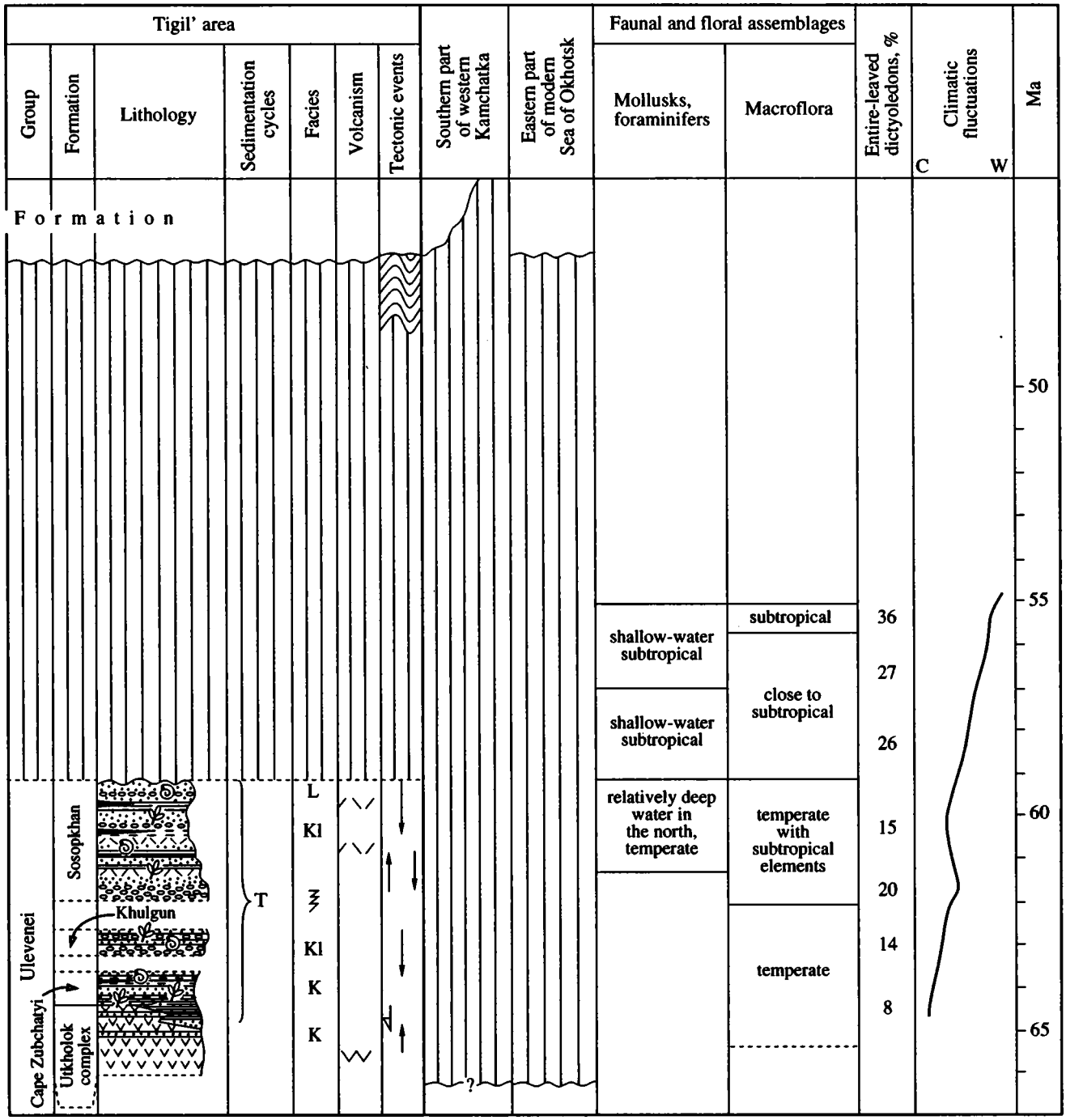


Fig. 4. (Contd).

the Lesnovskii uplift). It should be mentioned, however, that some lower Paleocene sections include rocks of volcanomictic and mixed composition, sometimes alternating with their subarkose varieties. This can be exemplified by the section of the Cape Zubchatyi Formation, which conformably but with insignificant hiatus overlies the Maastrichtian–Danian terrestrial volcanics of the Utkholokskii Peninsula. The lower boundary of volcanics is unexposed, but in the Khairyuzovaya River middle course, the Irunei Formation (presumably, the Maastrichtian in age) is conformably overlain by andesitic porphyrites, tuffbreccia, and volcanomictic sandstones lithologically very similar to Utkholok volcanics (*Geologiya SSSR*, 1964).

Comparing the dislocation patterns of Cretaceous and Paleogene sections, it should be noted that folding and fold–rupture dislocations in western Kamchatka, in

particular, and in Kamchatka as a whole are manifested very unevenly in both Cretaceous and Paleogene deposits, as well as in Cenozoic deposits in general. Irregular patterns of dislocations are registered both across and along the strike of major Kamchatka structures. For instance, systems of compressed inclined folds in the Upper Cretaceous Lesnovskii Formation are frequently (but not universally) present in the northern areas of the region, whereas in southern areas (e.g., Pravaya Napana river walls), folds in the probable stratigraphic analogues of this formation are simple and symmetric. Very irregular dislocations are characteristic of the Irunei (Maastrichtian) and Kirganik (Maastrichtian–Danian) formations in the southern part of the Sredinnyi Range of Kamchatka: the strained folding and plication in basins of the Icha and Bogdanovskaya rivers grade further southward (basins of Andrianovka, Evseika, and Zhupanka rivers) into large, frequently open, folds, symmetric or of slightly tilted in the eastern direction.

The same is also true for the lower Paleogene sequences of western Kamchatka. For instance, the highest dislocation degree is observed only within a relatively narrow band in the westernmost part of the region. Thrusts in this area with the displacement magnitude of several kilometers quickly attenuate along the strike, and systems of strained, frequently overturned (up to horizontal), folds are marked in axial parts by a distinct cleavage (similarly to folds in Cretaceous sequences). The strained folding is also registered in the axial part of the Tigil' uplift, where steep folds, subvertical rock attitude, and cleavaged conglomerate pebbles are observed in coarse psephites (Khulgun and Sosopkhan formations). Flank areas of the uplift are commonly characterized by simple symmetric and inclined folds and monoclines.

The listed data probably indicate the absence of general structural reorganization directly at the Cretaceous–Paleogene boundary in the western Kamchatka region. Some comparative materials on other structural zones contribute to solving this problem for all of Kamchatka.

In central Kamchatka (the southern part of the Srdeinnyi Range), the Maastrichtian Irunei Formation is conformably overlain by the sequence of subaerial–subaqueous volcanics (the Karganik Formation), which, in turn, are conformably overlain by terrigenous mainly volcanomictic deposits. According to radiolarian dating, the Kirganik Formation is considered to be Maastrichtian in age (Zinkevich *et al.*, 1993). Isotopic dating of gabbro–pyroxenite intrusions comagmatic with the Kirganik volcanics (Flerov and Koloskov, 1976) revealed their Paleogene age (dates obtained at the Laboratory of Isotopic Age, Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry, Russian Academy of Sciences, are as follows: 85.2 ∓ 3, 55 ∓ 3, 53 ∓ 1 Ma for biolite fractions and 56 ∓ 4 Ma for the whole-rock sample). These dates suggest that the Cretaceous–Paleogene boundary lies either within the Kirganik sequence or at its base, thus excluding, the presence of an unconformity at this level in central Kamchatka. At the same time, similarly to western Kamchatka, the sharp angular unconformity is registered in this region at the base of middle Eocene deposits overlying the Kirganik Formation.

The continuous Cretaceous–Paleogene section in the Eastern Kamchatka ranges was described as early as the 1960s (Shantser, 1966). In the 1980s, the Golubovskaya Formation from the lower part of this section in the Valaginskii Ridge was dated by radiolarians as late Maastrichtian in age. The formation is conformably overlain by volcanics of the Kitil'gin Formation, which, in turn, is conformably overlain by the terrigenous flyschoid Tal'nikovaya Formation with rare foraminifers of the Danian–Thanetian age. Thus, the continuous succession of Cretaceous and Paleogene deposits in eastern Kamchatka is beyond question. In eastern flanks of the Eastern Ranges (Valaginskii and Kumroch ridges), the facial analogue of the Tal'nikovaya Formation is represented by the Vetlovskaya Formation (Vetlovskaya complex), some fragments of which (usually in thrust sheets) yielded planktonic foraminifer assemblages of the Thanetian and early Ipresian ages. Cretaceous and lower Paleogene deposits of the Eastern Ridges are overlain with a sharp angular unconformity by the middle–late Eocene terrigenous sequences (Bakhteev *et al.*, 1994).

Thus, there are no grounds to assume the large-scale tectogenesis in the entire Kamchatka Peninsula at the time of the Cretaceous–Paleogene boundary, although breaks in deposition and intraformational hiatuses at this level cannot be completely excluded.

A similar situation is characteristic of the Cretaceous–Paleogene boundary in the North Pacific regions adjacent to Kamchatka. For instance, in the Olyutor zone and southern Koryak Highland (in sections of the Il'pinskii Peninsula and Mainy-Kakyine Ridge, respectively), Danian deposits conformably rest upon Maastrichtian layers. The presence of conglomerates and coarse psephites at this level most likely suggests rather sharp changes in hydrodynamics in the deposition basin, but does not exclude the possibility of discrete sedimentation and absence of some biostratigraphic zones in the sequence. Hiatuses without apparent angular unconformities are registered at this level in sections of Chukotka and Alaska (Burk, 1965), whereas, in Sakhalin, different researchers mention both the conformable relations between strata and local angular unconformities (Kovtunovich, 1969).

Nonetheless, the above-mentioned data do not mean that we are dealing with a universally continuous deposition. In many cases, we observe a substantially large hiatus and accumulation of thick conglomerate members; block movements accompanied by the formation of horst and graben structures that accumulated coastal-marine and continental deposits also cannot be disre-

garded. This is particularly characteristic of western Kamchatka, where an intense subaqueous–subaerial volcanism also occurred. Volcanism of this type during the late Maastrichtian–Danian was characteristic of the major part of the Kamchatka region. In westernmost areas located probably along the Okhotsk land margin, volcanism generally was of the calc-alkaline type, inherited afterward in the middle and late Eocene. In central Kamchatka, preponderant are volcanics of the absarokite–shoshonite association. According to some data, the eastern areas of the region (Eastern Ridges) host the island-arc tholeiites and calc-alkaline rocks, as well as such "exotic" (for the continental margin) rocks as meimechites, comatiites, and subalkalic picrites. In addition, mineral associations in some volcanics from these areas are sometimes close to kimberlites (Seliverstov and Grechin, 1979). All this permits an assumption that, during the Maastrichtian–Danian, volcanism of the region in question predominantly was subaerial and occurred on the land in the west and on a shallow sea archipelago in the east. Nonetheless, it is not inconceivable that the formation of three volcanic zones—western, central, and eastern all related to the development of deep faults of the northeastern strike—commenced at that time.

As follows from the aforesaid, the large-scale tectogenesis was probably not characteristic of the Cretaceous–Paleogene boundary that was marked only by such geological events as volcanism, block movements, and shallowing of the past sea basin with discrete (owing to submarine faults) deposition. It may be assumed that a system of closely interrelated sedimentation basins existed in the Kamchatka region during the Late Cretaceous–early Paleocene. In order to understand causes of the formation of various Paleocene formations in the region and estimate the type and age of the basement, on which these formations accumulated, we should consider some geological events and particular aspects of the Okhotsk Sea region geological evolution.

The Okhotsk land in the Late Cretaceous and Paleogene. In the south of the Sredinnyi Range of Kamchatka, there is an uplift of metamorphic, intensely dislocated rocks, some of which, namely the rocks of the Khimkin and Kheivan formations, are considered to be Cretaceous in age. In this very area (basins of the Oblukovina and Icha rivers), metamorphic rocks are overlain with a hiatus and sharp angular unconformity by the Barab Formation mainly composed of conglomerates and conglobreccia attributed, on the basis of paleofloral assemblages, to the middle–late Campanian. The upper portion of the formation, where continental facies are replaced by marine deposits, yielded the Campanian–Maastrichtian radiolarians (Shapiro *et al.*, 1986). The hiatus and angular unconformity are also registered in the Eastern Ridge (V.A. Seliverstov, pers. comm.).

Angular unconformities, hiatuses, and continental sequences are also recorded in other regions framing the North Pacific: in the Koryak Highland, Chukotka, Alaska, and Sakhalin (Burk, 1965; Herman and Lebedev, 1991). In addition, some researchers noted an unconformity at the base of the Maastrichtian section (Volobueva and Krasnyi, 1979) known as the so-called Maastrichtian neoautochton in the Koryak Highland. According to G.I. Zhegalova (pers. comm.), one of the boreholes drilled in the Kolpakovka area of western Kamchatka penetrated the rock sequence with Maastrichtian radiolarians resting on the intensely dislocated basement presumably of the Cretaceous age. On the other hand, no marine Campanian–Maastrichtian deposits are known along the northern and northwestern coasts of the Sea of Okhotsk, and the only terrestrial volcanic sequences of this age occur in the Okhotsk–Chukotka belt. The borehole drilled in the Magadan shelf recovered the middle Eocene deposits directly overlying the dislocated presumably Cretaceous deposits. These data evidence a considerable age gap between the basement and sedimentary cover of the Sea of Okhotsk that is also supported by many marine seismic profiles (Shainyan *et al.*, 1989), although the lower Paleogene continental (?) deposits occur in some areas.

The data discussed indicate that the Campanian is marked by regional angular unconformities related to major tectonic events. It seems probable that the Okhotsk land represented an uplifted fold zone at that time. Due to erosion, this zone provided abundant arkose and volcanomictic material to sedimentary basins formed to the east of it in the Campanian and evolved till late Paleocene–early Eocene. This assumption is supported by finds of abundant pebbles and boulders of volcanic rocks unknown amid Cretaceous deposits of Kamchatka and associated with clasts of metamorphic rocks in the Paleocene sequences of western Kamchatka. The intense erosion and denudation of the Okhotsk land proceeded till the middle Eocene, when the first large Cenozoic transgression occurred and, as a result, the sedimentary cover of the Okhotsk platform began to be formed. It should be stressed that we consider, following some researchers (Udintsev, 1981; and others), the larger part of the modern Sea of Okhotsk as representing a young Cenozoic platform with the Mesozoic folded basement that experienced, in the Miocene, a tectonic activation accompanied by the formation of deep-sea basins (including those of the oceanic type) and separate rises.

GEOLOGICAL HISTORY AND PALEOGEOGRAPHY OF WESTERN KAMCHATKA IN THE PALEOCENE

As was mentioned, the Cretaceous–Paleogene boundary in western Kamchatka is marked by subaerial and locally subaqueous calc-alkaline volcanism, although alkalic rocks (trachybasalts and trachyandesites) also occur in this zone. These early Paleogene vol-

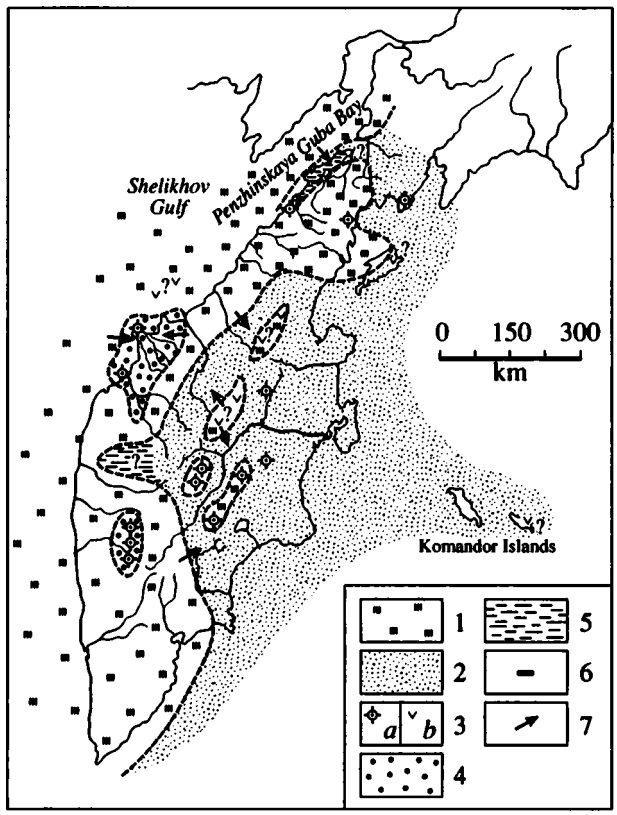


Fig. 5. Schematic paleogeographic map for the early Danian: (1) land; (2) sea; (3) proven (*a*) and assumed (*b*) volcanism occurrences; (4) continental sedimentatry basins; (5) sedimentation in basins with sluggish circulation; (6) coal accumulation; (7) direction of the clastic material transport.

canics represent a southern element of the western Kamchatka–Koryak volcanogenic belt that extends northward via the Koryak Highland to the Anadyr' depression. For comparison, this time interval in easterly areas of Kamchatka was marked by subaqueous–subaerial volcanism. As mentioned above, volcanoes were confined to certain areas of the region, probably, to three linear zones of the northeastern strike. In any case, the volcanic activity occurred under conditions of considerable shallowing of the past sea basin followed by appearance of volcanic and non-volcanic islands.

Thus, the Kamchatka Peninsula (except for its southern extremity) represented in the late Maastrichtian(?)–early Danian an area of the preponderant development of subaerial–subaqueous volcanism (Fig. 5), although data on the entire region are incomplete because its central part is nowadays covered with thick Miocene–Quaternary deposits, whereas its western part is submerged under the Sea of Okhotsk.

It seems probable that the barrier archipelago between western Kamchatka and easterly sea basins appeared exactly at that time. Simultaneously, in western Kamchatka and evidently along the Okhotsk land margin, a large system of linear faults of the northeastern strike started to be formed. Within the latter, the differentiated block movements resulted in the formation of grabens and horsts, and, most likely, in partial desiccation of the Western Kamchatka sea basin and in growth of separate rises composed of Cretaceous rocks. This can be exemplified by the occurrence of the late Danian–Thanetian Anadyrka Formation on one of such rises. Later, namely in the middle and late Eocene, subaerial volcanics were erupted in this zone of the inherited Maastrichtian–Danian volcanism. The early Paleogene (and Paleogene in general) volcanism in western Kamchtaka, as well as in the entire western Kamchatka–Koryak belt, was, probably, related to the formation of rift systems (or simply grabens). The position of the southern extremity of these systems near the continental margin (marginal–continental belt) was responsible for the compositional difference between volcanics of western Kamchatka and those of typical rift structures.

The western Kamchatka sea basin can be considered as a remnant one relative to the Late Cretaceous sedimentary basins. According to drilling data, it reached the modern Kolpakov River latitude (the southern part of western Kamchatka represented at that time a land composed by Cretaceous and older rocks, including their metamorphic varieties) in the south and the latitude of the Il'pinskii Island in the north. On the west, the basin was bounded by the intensely eroded Okhotsk land and on the east, most probably, by a chain of islands separated by intermittently existed straits. The most stable connection of this basin with easterly basins, both in its southern and northern parts, presumably existed in the late Danian–early Thanetian. For instance, in the northern areas, this is evidenced from the distribution of the typical marine Getkilnin Formation (Fig. 6). The terrigenous material was supplied from both the west (mainly arkose and subarkose material) and east, from a hypothetical archipelago most likely composed of Upper Cretaceous volcanogenic and terrigenous deposits. This can explain the mixed arkose–volcanomictic composition of terrigenous sequences frequent in the western Kamchatka sedimentary basin.

Mollusks and foraminifers penetrated into this basin from the east mainly via intermittently opened seaways. During the late Danian and Thanetian, marine conditions were particularly characteristic of the northern part of the basin, where the depth range sometimes corresponded to the lower subtidal–bathyal zone. This (northern) basin was relatively small and, probably, semiclosed. Sometimes, it was connected via shallow-water seaways with the southern part of the western Kamchatka basin. This assumption is supported by the presence of marine molluscan and foraminiferal assemblages in the Chemurnaut sequences, which are similar to those occurring in many areas of the North Pacific (including California) and even in the Atlantic and Thetys. The shallow-water (mainly sublittoral and,

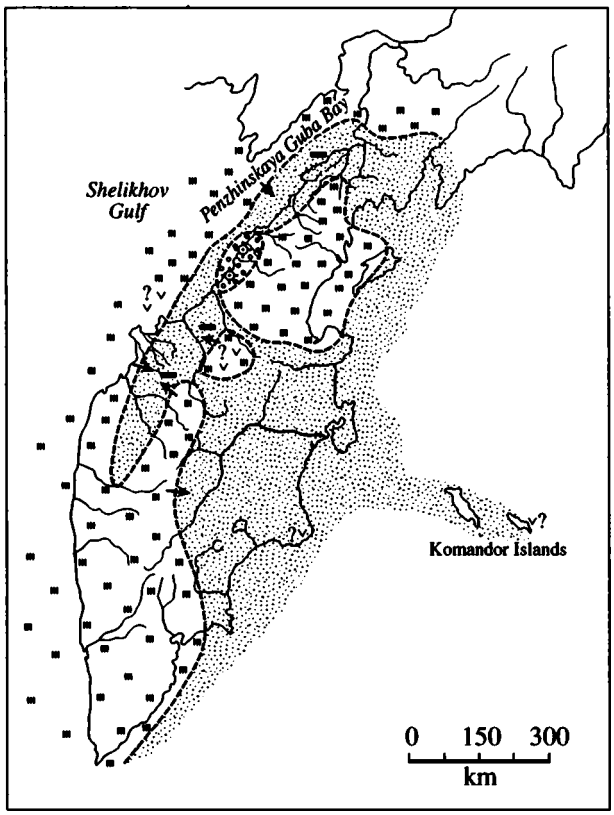


Fig. 6. Schematic paleogeographic map for the late Danian–early Thanetian (for legend see Fig. 5).

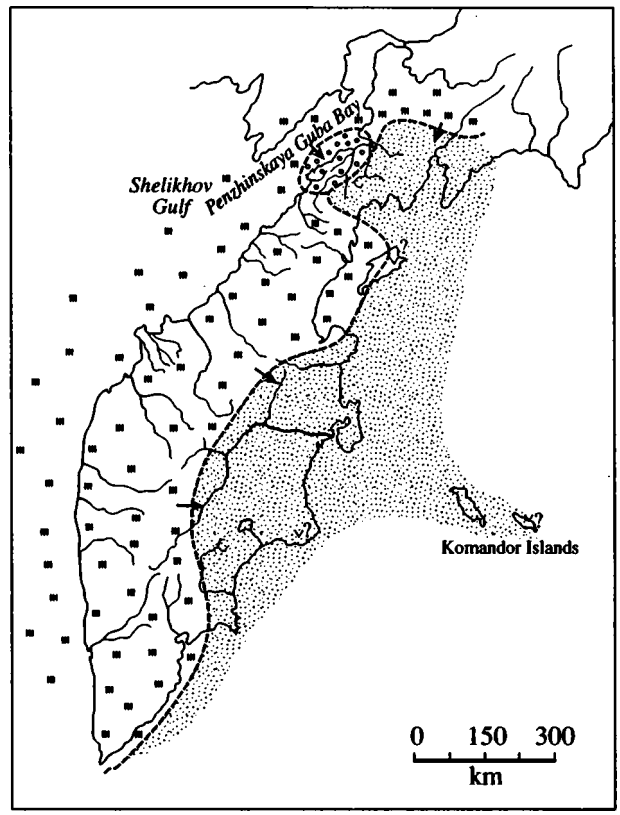


Fig. 7. Schematic paleogeographic map for the late Thanetian–early Ipresian (for legend see Fig. 5).

partly, bathyal) forms from these assemblages indicate a wide migration of biota within shelf areas of the Pacific during the early Paleogene.

There are, however, differences in the evolution of the southern and northern parts of the basin. Sedimentation in its southern part commenced slightly earlier than in the northern part and characterized a transgressive cycle preceding or partly coinciding with initiation of marine sedimentation in the northern part (the Getkilnin Formation). Thus, it may be assumed that the northern part of Kamchatka (from the Palana River on the south to the Pustaya River mouth on the north) represented in the early Danian a relatively uplifted block with the exposed Upper Cretaceous deposits. The area of the Tigil' uplift was located, on the contrary, on a relatively subsided block intermittently flooded from the east by a shallow sea (Fig. 6). In the late Danian (probably, in the earliest Thanetian), this area was sometimes a locus of acid volcanism, which is recorded in deposits of the Sosopkhan Formation. Volcanics of the southern part of the northern block (the Anadyrka Formation and others) are likely of the same age. The northern block was separated from the southern one probably along the fault of the northwestern strike that crosses Kamchatka, from the Palana River mouth on the west to the Ozernoi Peninsula on the east. Movements along this fault recommenced in the Eocene and Miocene. From our point of view, it represents the northeastern element of the fault system, the predecessor of the Aleutian island arc, which became active in the late Paleocene.

The sea basin within the northern block of western Kamchatka most probably appeared in the late Danian–Thanetian after a sharp reduction of the volcanic activity: no traces of volcanism are established in the marine Getkilnin Formation. A distinct regressive cycle with purely continental facies in the northern portion of the sequence (the Tkapravayam Formation and Evravayam Beds) can be outlined in this area (Fig. 7).

Inasmuch as no Thanetian marine deposits are observed in the southern block of western Kamchatka (they were probably eroded during the subsequent uplift), one can suggest the existence of an integrated sea basin in western Kamchatka at that time. It is also probable that somewhat earlier the southern block was involved in the uplifting movements and turned into the denudation area, whereas grabens accumulating thick continental sequences were locally formed in the northern block (Fig. 7).

In general, the lower Paleogene formations of western Kamchatka can be considered as molasses, commonly coal-bearing, and close to paralic. The presence of abundant flora-bearing interlayers with plant burial of a mainly autochtonous type and assemblages of

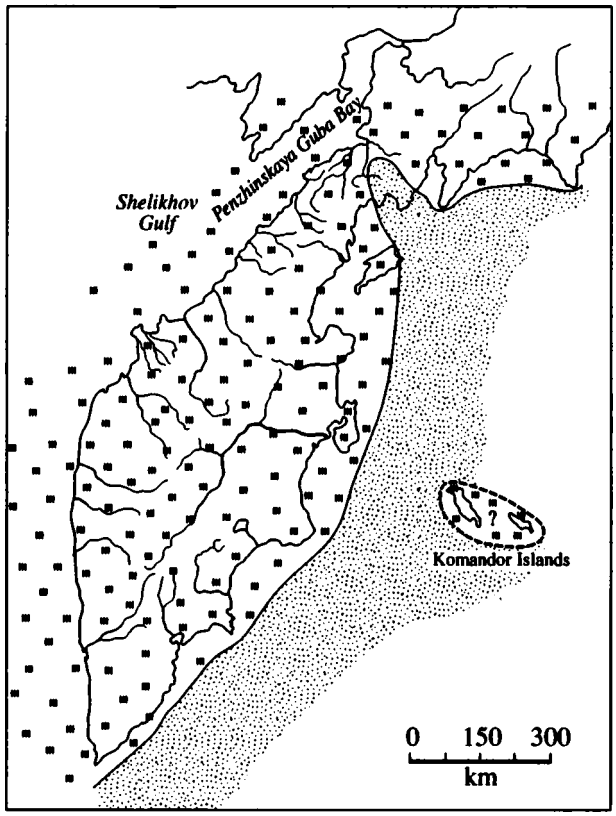


Fig. 8. Schematic paleogeographic map for the late–earliest middle Eocene (for legend see Fig. 5).

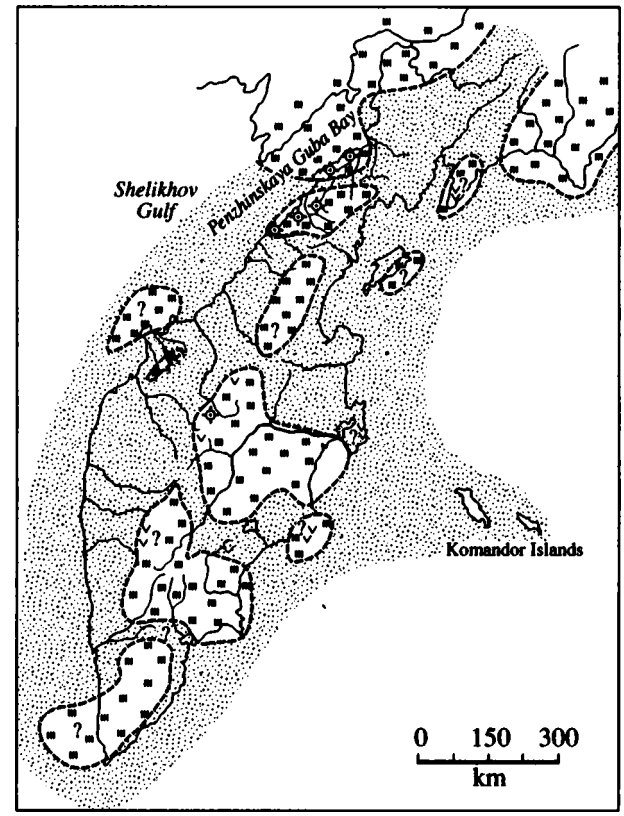


Fig. 9. Schematic paleogeographic map for the latest middle–earliest late Eocene (for legend see Fig. 5).

freshwater, euryhaline, and typical coastal-marine faunas, together with peculiar structural features of deposits (oblique, cross–wavy, and decussate bedding), indicate the sedimentation environments of a maritime plain and shallow shelf. At the same time, finds of typical slope vegetation suggest a sufficient relief differentiation in some areas of western Kamchatka (Budantsev, 1989). This is also supported by signs of avalanche sedimentation, which resulted in the formation of prograding facies particularly characteristic of the Tkapravayam Formation in the northern part of the region. Such a type of sedimentation was related to a rather fast formation of contrast structures (horsts and grabens) responsible for a sharp increase of clastic material mass transported by rivers to the deepest parts of the basin and for the accumulation of thick rock sequences. As for the more general paleogeograpic aspect, the analysis of paleofloras reveals a wide floral exchange between North America and East Asia, which commenced during the period of high land level in the Campanian and continued, probably with some breaks, till the beginning of the middle Eocene. Thus, one can suggest that the quite spacious Beringia land or shallow shelves with island archipelagoes existed in the Paleocene–early Eocene (Chelebaeva and Shantser, 1993).

As a result of the progressive uplifting, all of western Kamchatka turned into land by the end of the Thanetian, and the sea basin that existed there during the Paleocene became virtually closed. However, in the east, on the territory of the modern Kamchatka depression, Eastern Ranges, and eastern Kamchatka coast, the sedimentation continued. This is evident from the occurrence of upper Paleocene relatively shallow-water sequences recovered by drilling in the central Kamchatka depression (Berson *et al.*, 1981). On the other hand, the terrigenous subarkose flyschoids, which can be partly classed with turbidites of the continental slope and rise, were deposited in the region of the Eastern Ranges. Further eastward, they graded into facies of a relatively deep sea, where siliceous–calcareous mud accumulated and volcanism of the oceanic affinity (the Vetlovaya Formation) occurred.

Such a paleogeographic profile can be reconstructed, however, only for the central part of the region in the area between the Ozernoi Peninsula on the north and Shipunskii Peninsula on the east. It seems that this part of the region, western Kamchatka included, represented in the Thanetian (probably, at the beginning of the Ipresian) a continental margin resembling that of the passive type.

The tectonic activation of the continental margin related to general rising of the region at the beginning of the Eocene resulted in the formation of thrust-and-fold structures incorporating the upper Campanian–

Maastrichtian and Paleocene–lower Eocene formations (Fig. 8). The erosion and base-levelling were followed at the end of the middle Eocene by an extensive sea transgression over the significant part of Kamchatka and Okhotsk land of the Paleocene time (Fig. 9).

All these paleogeographic conclusions are drawn without regard to palinspastic reconstructions (because of the absence of data required), therefore we do not consider different geodynamic concepts of the continental margin evolution.

CONCLUSION

The study of lower Paleogene deposits of Kamchatka provided the following results.

(1) A new stratigraphic scheme proposed for the Paleocene deposits of western Kamchatka includes four regional stages (horizons within the Danian–Thanetian interval) and is used as a basis to compile a calendar of geological events.

(2) A series of schematic paleogeographic maps is compiled for different time slices of the early Paleogene (Danian–Thanetian); these maps demonstrate a degradation of sea basins and expansion of land area in the eastern part of western Kamchatka.

(3) It is established that paleoclimatic events in Kamchatka correlate with those in the North Pacific.

(4) It is inferred that the Cretaceous–Paleogene boundary in the western Kamchatka region is marked only by block movements and subaerial volcanism, but not by major structural reorganization and changes in paleogeography. The Paleocene sedimentary basins can be viewed as remnant structures relative to basins which existed in the Maastrichtian.

(5) It is established that the formation of the Okhotsk land commenced in the late Campanian to be a denudation area till the middle Eocene. The formation of the sedimentary cover in the Sea of Okhotsk was related to the first Paleogene (the second half of the middle Eocene) transgression that was intense and followed the stage of tectonogenesis and erosion.

(6) Kamchatka of the Paleocene time can be considered as representing the continental margin resembling that of the passive type.

ACKNOWLEDGMENTS

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Oligocene Calcareous Plankton from the Kronotskii Peninsula, Eastern Kamchatka

S. I. Stupin*, M. Ya. Serova, E. A. Shcherbinina*, D. I. Vitukhin*, N. V. Tsukanov***, and A. S. Pachkalov*****

** Geological Institute (GIN), Russian Academy of Science, Pyzhevskii per. 7, Moscow, 109017 Russia*
*** Paleontological Institute, Russian Academy of Sciences, Profsoyuznaya ul. 123, Moscow, 117321 Russia*
**** Institute of Oceanology, Russian Academy of Sciences, ul. Krasikova 23, Moscow, 117581 Russia*
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Abstract—Planktonic foraminifers and calcareous nannoplankton were first recorded along with benthic foraminifers in the Upper Paleogene section of the Kronotskii Peninsula, eastern Kamchatka. The fauna permits the assignment of the enclosing sediments to the *Globigerina ampliapertura* Zone of the foraminifer zonation of Bolli and Saunders, (1985) and to the nannoplankton zones CP17-CP18 of the Lower Oligocene. The sporadic occurrence of calcareous plankton in Oligocene sediments of high latitudes was most likely a consequence of the World Ocean level rise and relative climatic warming.

Key words: *Far East, eastern Kamchatka, foraminifers, nannoplankton, zone, Oligocene.*

INTRODUCTION

The correlation of regional stratigraphic subdivisions with standard zonations of different planktonic microfossils is an urgent problem in Cenozoic geology and stratigraphy in the Far East of Russia and, specifically, in eastern Kamchatka. The correlation is generally possible in two ways. The first one is a correlation of subdivisions distinguished by benthic fauna, namely, by mollusks and benthic foraminifers, with those of Japan or North America, where plankton occurs along with benthic fossils. The second way is an active search for plankton remains in the Far East sections to correlate them with the oceanic zonation. In the former case, the estimated age of deposits can be changed according to the verified dates of corresponding stratigraphic subdivisions in Japan or America. The disadvantage of the second way is the relatively scarce occurrence of plankton, especially of calcareous microfossils, in Cenozoic sediments of the Far East of Russia. The rarity of plankton remains is unfavorable for a detailed stratigraphic subdivision of sediments in contrast to the benthic fauna; however, the planktonic fossils may reveal some solid stratigraphic markers. The aim of this paper is to outline one such a marker, the *Globigerina ampliapertura* Zone, in the Cenozoic section of the Kronotskii Peninsula, eastern Kamchatka.

In 1994, Tsukanov and Pachkalov studied the Cenozoic deposits of the Tyushevskii Trough in the Rakitinskaya River basin, the Kronotskii Peninsula of eastern Kamchatka (Fig. 1). They collected samples for the microfaunal analysis and described the lithostratigraphy of the studied sections. Additionally they used the samples taken by V.N. Sinel'nikova (GIN RAS) from the sediments of the Rakitinskaya Formation in 1989 during field work with M.E. Boyarinova (Kamchatgeologiya). Other authors of the paper performed the following investigations. Serova and Stupin studied foraminifers, Shcherbinina studied nannoplankton, and Vitukhin attempted to extract radiolarian and diatom remains and considered the history of investigation of the Rakitinskaya Formation and its age estimates.

BRIEF GEOLOGICAL DESCRIPTION AND HISTORY OF INVESTIGATION OF THE RAKITINSKAYA FORMATION

The Rakitinskaya Formation is exposed within the Tyushevskii Trough that extends northeastward as a narrow, about 60-80 km wide, structure, from the Shipunskii Peninsula on the south to the Kamchatskii Cape on the north. It separates the eastern Kamchatka uplifts from the peninsulas of eastern Kamchatka and is composed of terrigenous and tuffaceous–terrigenous Cenozoic deposits. The trough is asymmetric; the thick sediments of the western flank are strongly deformed and experienced thrusting, whereas on the eastern flank the deposts of the same age compose a gentle monocline disturbed by faults.

Shtempel' and his colleagues (1935), the first researchers of the Kronotskii Peninsula, distinguished upward from the base the Kronotskii, Bogachevka, and Tyushevka formations ranging in age from the Paleogene to the Miocene. Later investigations of the region demonstrated that, because of the lithologic variability of rocks, their strong deformation, and poor paleonto-

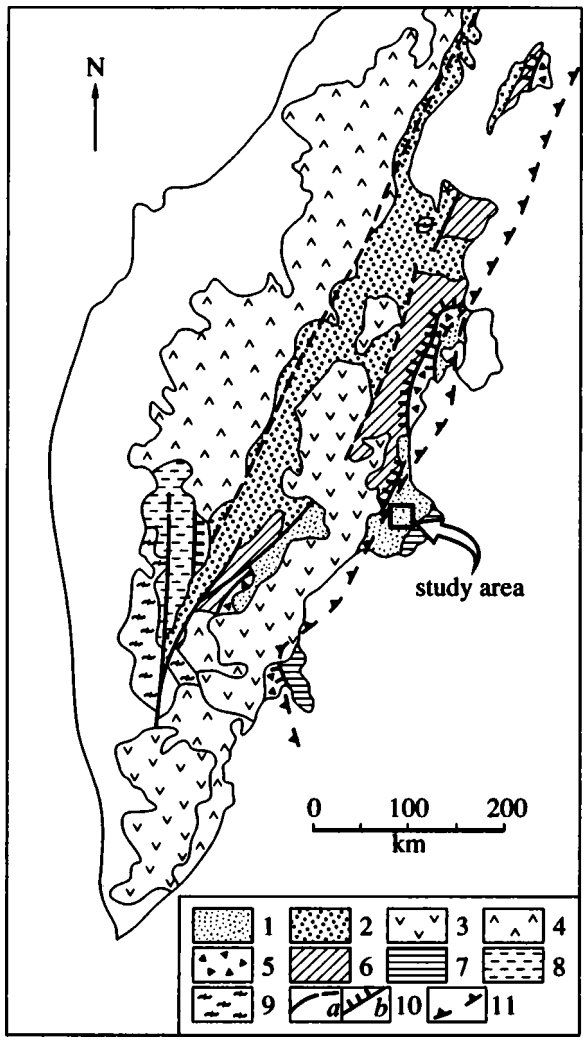


Fig. 1. Tectonic scheme of Kamchatka (after Zinkevich and Tsukanov, 1992): (1, 2) Cenozoic terrigenous deposits of the (1) Tyushevskii Trough and (2) Central Kamchatka graben; (3, 4) volcanogenic rocks of (3) East Kamchatka belt (N_2–Q) and (4) Central Kamchatka belt (P_3–N_1); (5) Vetlovka complex (K_2–P_2^2); (6) volcanogenic, volcanogenic–sedimentary, and terrigenous deposits of the Eastern Kamchatka zone (K_2–P_1); (7) complexly deformed terranes of the eastern peninsulas; (8) volcanogenic–sedimentary and terrigenous sediments of the Central Kamchatka zone; (9) metamorphic rocks (Mz-Kz); (10) proven and assumed faults (*a*) and thrust faults (*b*); (11) Vetlovka collision suture.

logic remains, the suggested stratigraphic scheme is discrepant and needs a revision.

The Rakitinskaya Formation was first identified by Pleshakov and Nesvit (1958) as a unit (subformation) of the Tyushevka Formation. Subsequently, it was classified as a formation in the Tyushevka Group. In the Kronotskii region, these authors distinguished an upward succession of the Bogachevka, Ivanova, and Tyushevka formations divided into a series of subformations. According to Pleshakov and Nesvit, the Rakitinskaya Subformation is represented by flyschoid interlayering of tuffaceous opoka-like shales and sandstones about 400 m thick. Its stratotype was not indicated,[1] and it was only noted that deposits of the subformation are distributed along the axis of the Rakitinskaya syncline that extends northeastward along the isthmus of the Kronotskii Peninsula from the Ol'ga Bay to the Malaya Chazhma River mouth. According to Pleshakov and Nesvit (1958), the Rakitinskaya Subformation crowns the Miocene section in the region and is an analogue of the Goryachie Klyuchi Formation distinguished previously along the Tyushevka River and dated back to the Pliocene (Dvali, 1955). However, it was mentioned in the same paper (Pleshakov and Nesvit, 1958) that A.P. Il'ina and I.G. Pronina, who studied the mollusk fauna of the Rakitinskaya Subformation, inferred its middle (latest) to the late Miocene age. A discrepancy between the age estimates of the unit was not explained in the paper.

Subsequently, the age of the Rakitinskaya Formation was considered as ranging from the early Miocene to the Pliocene. At the Okha Stratigraphic Conference, its age interval was adopted as the late Miocene–early Pliocene (*Unifitsirovannye* ..., 1961). Il'ina (1963) defined the age of the formation as the late Miocene, and Pronina (1969) as the middle Miocene. Later on, Pronina and Berson (1978) assigned the Rakitinskaya Formation to the lower Miocene.

Arsanov (1978), who studied the geology of the Kronotskii region in 1963–1966, noted that the identification of the Rakitinskaya Formation with the top unit of the Miocene section was erroneous. He demonstrated that the Rakitinskaya Formation unconformably overlies the basalts of the Kronotskii Group and is in turn discordantly overlain by sediments of the Tyushevka Group. The unit was subdivided into three subformations: the lower one of interlayered sandstones, siltstones, and gravelstones; the middle one of conglomerates and sandstones crowned with spongolites incorporating opal cherts, whose beds are intercalated with glauconite sandstone layers bearing a nodular phosphorite interlayer; and the upper subformation mainly composed of sandstones and siltstones. Arsanov estimated the maximum thickness of the Rakitinskaya Formation to be 235 m.

According to his data, a single section of all three subformations is exposed at the right bank of the Tyushevka River near the Volch'ya River mouth and corresponds to the Goryachie Klyuchi Formation after D'yakov and Dvali (1955). The section is mainly composed of clastic rocks: boulder conglomerates, gravely sandstones, sandstones, and coquina. To the north, in the Rakitinskaya River basin, the formation is represented by sandstones and siltstones bearing the coquina, glauconite sandstone, and spongolite layers,

[1] Pronina (1969) stated that the stratotype of the Rakitinskaya Formation is situated in the upper reaches of the Rakitinskaya River without further details of its location.

with the boulder and pebble conglomerate bed at the base of the section.

Serova identified foraminifers of the Rakitinskaya Formation from the Arsanov's collection as representing an assemblage of the Oligocene age. On the basis of mollusks, Arsanov correlated the Rakitinskaya Formation with the Machigar Formation of northern Sakhalin and with the Alugivayam Formation of the southern Koryak Highland (Il'pinskii Peninsula), both of the Oligocene age as well. At the Second Interdepartmental Stratigraphic Conference of 1974, the Rakitinskaya Formation was included in the Alugivayam Horizon referred to the Oligocene (*Resheniya* ..., 1982). However, soon afterwards, the age of the formation was revised by Runeva (1979), who studied the radiolarian fauna from the Rakitinskaya Formation and dated it as middle Miocene in age. The tendency to rejuvenate the formation age is valid up to present. For instance, in the geologic report of Boyarinova, who studied recently the geology of the Kronotskii Peninsula, the formation is attributed to the middle Miocene. Based on the mollusk assemblages, Sinel'nikova (personal communication) is inclined to correlate it with the lower Miocene Kuluven Formation of the Tochilino section in western Kamchatka. Shapiro *et al.* (1996) correlated the Rakitinskaya Formation with the basal unit of the Kavran Group of western Kamchatka corresponding to the middle Miocene. Thus, there is no general agreement about the age of the Rakitinskaya Formation.

DESCRIPTION OF THE SECTIONS

Sections of the Rakitinskaya Formation were studied in 1994 by Tsukanov and Pachkalov in the middle reaches of the Rakitinskaya River (on the left bank) and in the area of the Vodopadnyi and a nameless creeks, the latter located one kilometer to the north of the former (Fig. 2).

In the middle reaches of the Vodopadnyi Creek, basalts of the Kronotskii Group are overlain by the following members (upward from the base, Fig. 3a):

(1) Medium- to fine-pebble (in the upper beds) conglomerates with reworked clastics of basic and intermediate extrusive rocks. The thickness is 1.5–2 m.

(2) Light gray medium-grained sandstones bearing fine pebbles of extrusive rocks and molluscan shell fragments. The thickness is 1.5 m.

(3) Pink gray coquina composed of 80% broken mollusk shells cemented by sandy or silty material. The thickness is 1.5 m.

(4) Slightly laminated opokas (Samples 18/18 and 18/19) enclosing foraminifers *Globigerina ampliapertura* Bolli, *G. ciperoensis angustiumbilicata* Bolli, *G. praebulloides* Blow subsp. indet., *Globocassidulina subglobosa* (Brady), *Islandiella excavata* (Volosh.), *Cibicidoides pulensis* Ten Dam et Reinhold, *C. tenellus* (Reuss), *Melonis shimokinensis* (Asano), *Bolivina marginata adeloidana* Cush. et Kleinpell, *Angulogerina gracilis germanica* Cush. et Edwards, *A. gracilis tenuistriata* (Reuss), and *Pullenia bulloides* (d'Orb.). The thickness is 1.5 m.

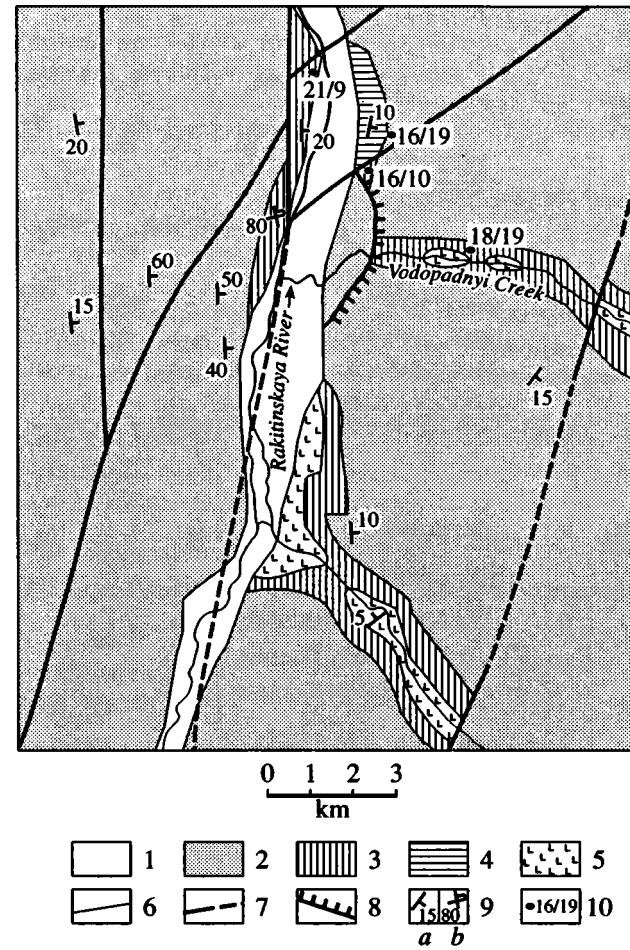


Fig. 2. Geologic scheme of the middle reaches of the Rakitinskaya River (after M.E. Boyarinova): (1) Quaternary alluvial deposits; (2) undivided Tyushevka Group; (3) Rakitinskaya Formation; (4) undivided Bogachevka Group; (5) undivided Kronotskii Group; (6) proven and assumed geological boundaries; (7) flaults; (8) thrust faults; (9) bed attitude (*a*) normal and (*b*) overturned; (10) site of samples bearing microfossils.

(5) Grayish green gravelstones with pebbles up to 1-2 mm in size. The thickness is 2 m.

(6) Dense tuffstones with the carbonate cement red on the weathered surface (Sample 18/21). The thickness is 15 m.

(7) Greenish, dense, poorly sorted sandstones (Sample 18/22). The thickness is 4 m.

(8) Loose sandstones of variable grain size. The thickness is 10 m.

In the section on the nameless creek, the Rakitinskaya Formation overlies the erosion surface of flyschoid–tuffaceous terrigenous sediments of the

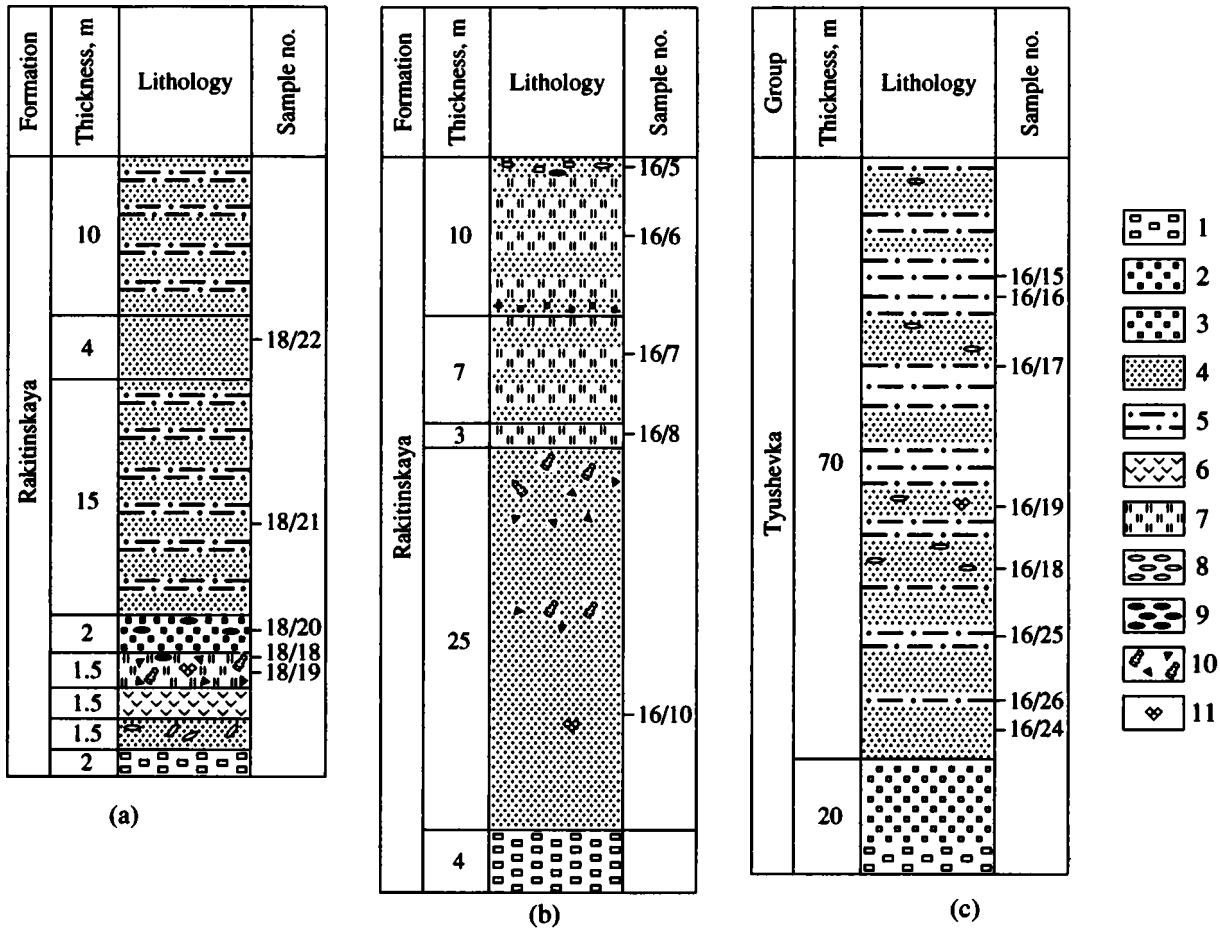


Fig. 3. Studied sections of the Rakitinskaya Formation and Tyushevka Group: (a) section of the Vodopadnyi Creek area (point 18/19 in Fig. 2), (b) section of the adjacent creek area (point 16/10), (c) section on the neighboring gulch (point 16/19); (1, 2) conglomerates (1) coarse pebble and (2) fine pebble; (3) gravelstones; (4) sandstones; (5) siltstones; (6) coquina; (7) opoka; (8) pebble; (9) carbonate concretions; (10) individual molluscan shells; (11) calcareous plankton.

Bogachevka (?) Group.[2] The formation is composed here of the following members (upward from the base, Fig. 3b):

(1) Fine-pebble conglomerates grading upward into gravelstones. The thickness is 4 m.

(2) Light gray to pinkish sandstones bearing dispersed molluscan fragments (Sample 16/10). A single nannoplankton species *Dictyococcites bisectus* (Hay, Mohler et Wade) is identified. The thickness is 25 m.

(3) Dense pinkish opoka bed with numerous molluscan fragments (Sample 16/8). The thickness is 3 m.

(4) Light gray opoka bed containing sandy admixture and interlayers of grey massive sandstones (Sample 16/7). The thickness is 7 m.

(5) Irregular interlayering of gray medium-grained massive sandstones and opokas bearing scarce pebbles of extrusive rocks and carbonate concretions (Samples 16/5 and 16/6). The thickness is 10 m.

A little northward of the nameless creek, in the adjacent gulch, the Rakitinskaya Formation is missing from the section, and turbidites of the Bogachevka (?) Group are overlain by sediments referred by Tsukanov and Pachkalov to the Tyushevka Group (Fig. 3c). At the base, the sediments are composed of fine-pebble conglomerates and gravelstones 20 m thick. The basal beds are overlain by a coarse rhythmic member of fine-grained feldspathic graywacke (Sample 16/24) with rare thin, up to 5 cm, silty graywacke beds (Samples 16/26 and 16/25). The rocks grade upward into sandy nodular lumpy siltstones (Samples 16/17, 16/16, and 16/15) bearing beds of fine-grained sandy graywacke and opoka (Sample 16/19). The opoka beds contain *Globocassidulina subglobosa* (Brady), *Cibicidoides tenellus* (Reuss), *Gavelinella californiensis* (Cush. et Hobson), *?Alabamina tangentialis* (Clodius), *Melonis pompilioides* (Fichtel et Moll), *Bolivina marginata adeloidana* Cush. et Kleinpell, and *B. fastigia* Cush. In addition, scarce reworked Cretaceous planktonic fora-

[2] Tsukanov and Pachkalov attributed these deposits to the Bogachevka Group with certain reservations. They differ from the rocks of the Rakitinskaya Formation and Tyushevka Group by a greater degree of lithification and deformation.

minifers *Globotruncana* sp. indet. and *Heterohelix* sp. indet. are recovered. They associate with the following nannoplankton forms: *Coccolithus pelagicus* (Wallich), *Cyclicargolithus floridanus* (Roth et Hay), *C. abisectus* (Muller), and *Reticulofenestra* cf. *minuta* Roth. The sandstones are characterized by graded bedding and yield the concretions of clayey-calcareous siltstones (Sample 16/18). The member as a whole is of a rhythmic structure, its siltstone and sandstone layers are 30-40 cm and 5-10 cm thick, correspondingly. The total thickness is 70 m.

Separated outcrops of the Rakitinskaya Formation occur at the left bank of the Rakitinskaya River valley, downstream of the Vodopadnyi Creek mouth. The sediments similar in lithologic composition with those exposed along the Vodopadnyi Creek were observed here. The fine-grained sandstones bearing siltstone and opoka lenses, and dispersed molluscan shell fragments are most common (Sample 21/9 collected from opoka beds two kilometers downstream of the Vodopadnyi Creek mouth; see Fig. 2). The opoka beds yielded *Globigerina ampliapertura* Bolli, *G. ciperoensis ciperoensis* Bolli, *G. praebulloides praebulloides* Blow, *Globorotalia opima nana* Bolli, *Fissurina soldanii* Seguenza, *Obliquina* cf. *hexagona* (Williamson), *O. semistriata* (Williamson), *O. substriata* (Williamson), *Globocassidulina subglobosa* (Brady), *Cibicidoides* aff. *elmaensis* Rau, *C. lopjanicus* Mjatl., *Gavelinella californiensis* (Cush. et Hobson), *Discorbis* aff. *globularis* (d'Orb.), *Valvulineria* aff. *willapaensis* Rau, *Bolivina marginata adeloidana* Cush. et Kleinpell, *Angulogerina gracilis germanica* Cush. et Edwards, *Islandiella excavata* (Volosh.), *Uvigerina* cf. *galloway blakeleyensis* Fulmer, and *Pullenia bulloides* (d'Orb.). Nannoplankton is represented by *Cyclicargolithus floridanus* (Roth et Hay) and *Reticulofenestra* cf. *minuta* Roth.

The sediments of the Rakitinskaya Formation and Tyushevka Group as a whole are slightly deformed and dip at angles of 10–15°. The intense degree of deformations was observed only close to faults. For example, the beds of the formation are overturned at the left bank of the Rakitinskaya River and in the nameless creek area we recorded the rock foliation near the faults and small folds of the northwest vergence.

INVESTIGATION METHODS

The studied thin-sections opokas indicated that the rocks are largely composed of recrystallized diatom, radiolarian, and sponge remains constituting 50–70% of the total detritus. Sponge spicules are far better preserved than the siliceous plankton. It is possible that, in some cases, Arsanov (1978) described opokas as spongolites (our data show that proportions between siliceous plankton and sponges in opokas are roughly equal). We failed to extract distinguishable diatom and radiolarian remains from these rocks and concretions.

In the sections described above, the opoka beds rich in biogenic carbonate constituting about 30% of the total detrital components and incorporating the well-preserved foraminifer shells and nannoplankton (Samples 18/19, 16/10, 16/19, and 21/9) are most interesting.

Because of the great hardness of opokas, the traditional maceration method, which was used for other samples, was inappropriate for extraction of foraminifers. They were extracted by processing samples with the concentrated hydrofluoric acid (HF) to substitute their shell carbonate by fluorite. The texture and structure of shells remain preserved after this treatment. Nannoplankton was extracted by the standard procedure and examined under the microscope.

Remains of planktonic and benthic foraminifers and nannoplankton were recorded in four samples (18/19, 16/10, 16/19, and 21/9) from collection of Tsukanov and Pachkalov and in nine samples (1031g/1, 1031/4, 1035/12, 1036/2, 1036/4, 1037/3, 99a, 103, and 107) from collection of Sinel'nikova.

ANALYSIS OF FORAMINIFER AND NANNOPLANKTON ASSEMBLAGES AND AGE ESTIMATES

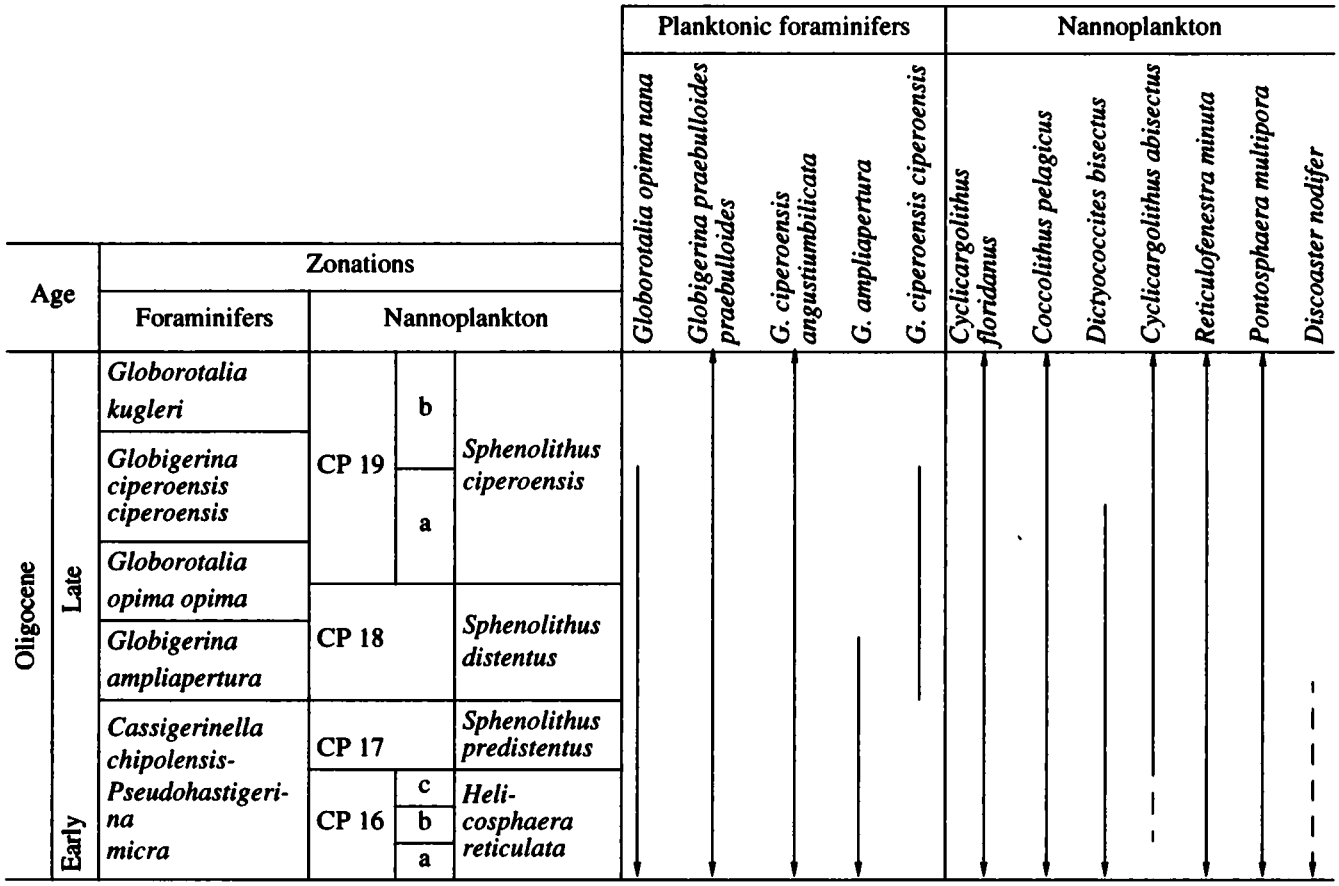
Before we turn to the discussion of the encountered assemblages, we should note that the examined planktonic foraminifers have some morphological peculiarities characteristic of the Cenozoic plankton from high latitudes (Asano and Hatai, 1967; Serova, 1978a; Krasheninnikov *et al.*, 1988). Globigerinas from the studied assemblage, like their specimens from the Paleocene–Eocene sediments of the northwestern Pacific, have not globular, but compressed chambers more tightly arranged as compared with the same species from the tropical and subtropical regions. They also have a narrower umbilicus and a less open aperture.

The joint occurrence of planktonic foraminifers *Globigerina ampliapertura* Bolli and *G. ciperoensis ciperoensis* Bolli indicates that the enclosing sediments correspond to the *Globigerina ampliapertura* Zone (the top of the Lower Oligocene) or, most likely, to a part of this zone (Table).

The recorded benthic foraminifers represent an integral assemblage. Changes in their taxonomic composition recorded in certain samples seem to be a result of different bionomic environments.

Many of the constituent benthic species are widespread in the Oligocene deposits of the Far East. *Gavelinella californiensis* (Cush. et Hobson) is known from the Amanina Formation and *Fissurina soldanii* Seguenza from the Utkholok Formation of western Kamchatka (*Atlas fauny* ..., 1984). *Globocassidulina subglobosa* (Brady) is recorded in the Amanina and Gakkh formations of western Kamchatka (*Atlas fauny* ..., 1984), in the Gennoishi Formation of western Sakhalin (Serova, 1985), and in the Mutnaya and Pilenga forma-

Stratigraphic ranges of calcareous plankton species from the Oligocene sediments of the Kronotskii Peninsula, eastern Kamchatka (after Bolli and Saunders, 1985; Perch-Nielsen, 1985)



tions of eastern Sakhalin. In the last case, this species coexist with *Cibicidoides lopjanicus* Mjatl. (Mitrofanova and Melent'eva, 1991). *Melonis pompilioides* (Fichtel et Moll) is known from the top of the upper Eocene Gailkhavilanvayam Formation of the Il'pinskii Peninsula (*Opornyi razrez* ..., 1994); from the top of the Ionaivayam Formation of the Koryak uplift (Serova *et al.*, 1977); from the Oligocene?–lower Miocene Bora Formation of eastern Sakhalin (Mitrofanova, 1984); and from the Poronai (upper Eocene–lower Oligocene) and Momijiyama (Oligocene) formations of Hokkaido, Japan, (Kaiho, 1984a, 1984b). *Melonis shimokinensis* (Asano) occurs in the Oligocene-middle Miocene sediments of western Kamchatka (Serova, 1978b; *Atlas fauny* ..., 1984), in the Il'khatunvayam Formation of the Karaginskii Island (Serova, 1978b), at the top of the Gailkhavilanvayam and in the Alugivayam Formation of the Il'pinskii Peninsula (*Opornyi razrez* ..., 1994), in the lower part of the Oligocene?–lower Miocene Kholmsk Formation of southeastern Sakhalin (*Opornyi razrez* ..., 1992), in the Mutnaya and Pilenga formations of eastern Sakhalin (Mitrofanova and Melent'eva, 1991), and in the Gennoishi Formation of western Sakhalin (Serova, 1985). *Islandiella exavata* (Volosh.) was found in the Oligocene–Miocene deposits of western Kamchatka (Serova, 1978b; *Atlas fauny* ..., 1984) and in the lower part of the Bora Formation of eastern Sakhalin (Mitrofanova, 1984; Mitrofanova and Melent'eva, 1991).

According to nannoplankton, the enclosing sediments are most likely not older than the early Oligocene, as the recorded *Cyclicargolithus abisectus* (Muller) appeared at that time.

Let us now consider the micropaleontological characteristic of the samples collected by Sinel'nikova in the same region, mainly along the right-hand tributaries of the Rakitinskaya River in its middle courses, the Krivoi and Borodavkin creeks. We should note that the microfauna was extracted from the samples collected for the malacologic analysis. The rocks are chiefly represented by sandstones and sandy siltstones bearing numerous mollusk remains.

All samples yielded an assemblage of benthic foraminifers and calcareous nannoplankton. Benthic foraminifers are represented by *Fissurina laevigata laevigata* Reuss, *Sigmomorphina* cf. *vaughani* Cush. et Ozawa, *Buccella* sp., *B.* ex gr. *subconica* Budasheva, *Cibicidoides tenellus* (Reuss), *C. lopjanicus* Mjatl., *Cibicides* cf. *lobatulus* (Walker et Jakob), *C.* ex gr. *celebrus* Bandy, *Melonis* cf. *shimokinensis* (Asano),

M. affine (Reuss), *Bolivina marginata adeloidana* Cush. et Kleinpell, *Cassidulina galvinensis* Cush. et Frizzell, *C. depressa* Asano et Nakomura, *C. menneri* Serova, *C.* sp., *Globocassidulina globosa* (Hantken), *G. subglobosa* (Brady), *Cassidulinoides howei* Cush., C. sp., *Islandiella exavata* (Volosh.), *I.* sp., and *Pullenia* ex gr. *salisburyi* R.E. et K.C. Stewart. Additionally, we recorded scarce planktonic foraminifers, whose taxa are undeterminable because of poor preservation.

This assemblage certainly resembles the previous one, but differs in a less diverse composition, occurrence of *Buccella*, and predominance of Cassidulinidae. This is most likely a result of sediment accumulation in the somewhat different settings of the outer subtidal or upper bathyal zone.

Some benthic foraminifers from samples of Sinel'nikova have wide stratigraphic ranges within the Eocene-Miocene interval. However, the forms known from the Upper Paleogene sediments predominate. For instance, *Sigmomorphina vaughani* Cush. et Ozawa and *Cibicidoides celebrus* Bandy occur in the Eocene deposits of North America; and in the Gakkh Formation (Oligocene) of western Kamchatka. *Cassidulina menneri* Serova was also described from the last locality (*Atlas fauny* ..., 1984). *Cassidulina galvinensis* Cush. et Frizzell was recorded in the Eocene and Oligocene sediments of North America; in the Kovachina (Eocene), Amanina, and Gakkh (Oligocene) formations of western Kamchatka; in the Alugivayam Formation (Oligocene) of eastern Kamchatka; and in the Krasnopol'e (Eocene) and Arakai (Oligocene) formations of Sakhalin (*Atlas fauny* ..., 1984; *Opornyi razrez* ..., 1994). *Globocassidulina globosa* (Hantken) was found in the Eocene-Oligocene beds of North America; in the Kovachina, Snatol (Eocene), and Amanina (Oligocene) formations of western Kamchatka; and in the Kholmsk (Oligocene?-lower Miocene) Formation of Sakhalin (*Atlas fauny* ..., 1984). *Melonis affine* (Reuss) is known from the Eocene of North America and the Amanina Formation of western Kamchatka (*Atlas fauny* ..., 1984). *Cassidulinoides howei* Cush. occurs in the upper part of the Poronai Formation (upper Eocene-Oligocene) and in the Nuibetzu Formation (Oligocene) of Hokkaido (Kaiho, 1984a, b). The available data on distribution of benthic foraminifers suggest the Oligocene age of enclosing deposits.

The following species of calcareous nannoplankton were identified in the same samples: *Reticulofenestra minuta* Roth, *Dictyococcites bisectus* (Hay, Mohler et Wade), *Cyclicargolithus floridanus* (Roth et Hay), *Pontosphaera multipora* (Kamptner), *Discoaster nodifer* (Bramlett et Riedel), and *Coccolithus pelagicus* (Wallich). This assemblage testifies to the fact that the sediments are not younger than the late early Oligocene, as the last occurrence level of *Discoaster nodifer* (Bramlett et Riedel) was recorded in the lower part of zone CP18, (*Sphenolithus distentus* zone (in table). Therefore, the joint occurrence of *Cyclicargolithus abisectus* (Muller) and *Discoaster nodifer* (Bramlett et Riedel) found in collections of Tsukanov and Pachkalov and Sinel'nikova, correspondingly, permits us to attribute the enclosing sediments to the lower Oligocene quite concordantly with data on planktonic foraminifers.

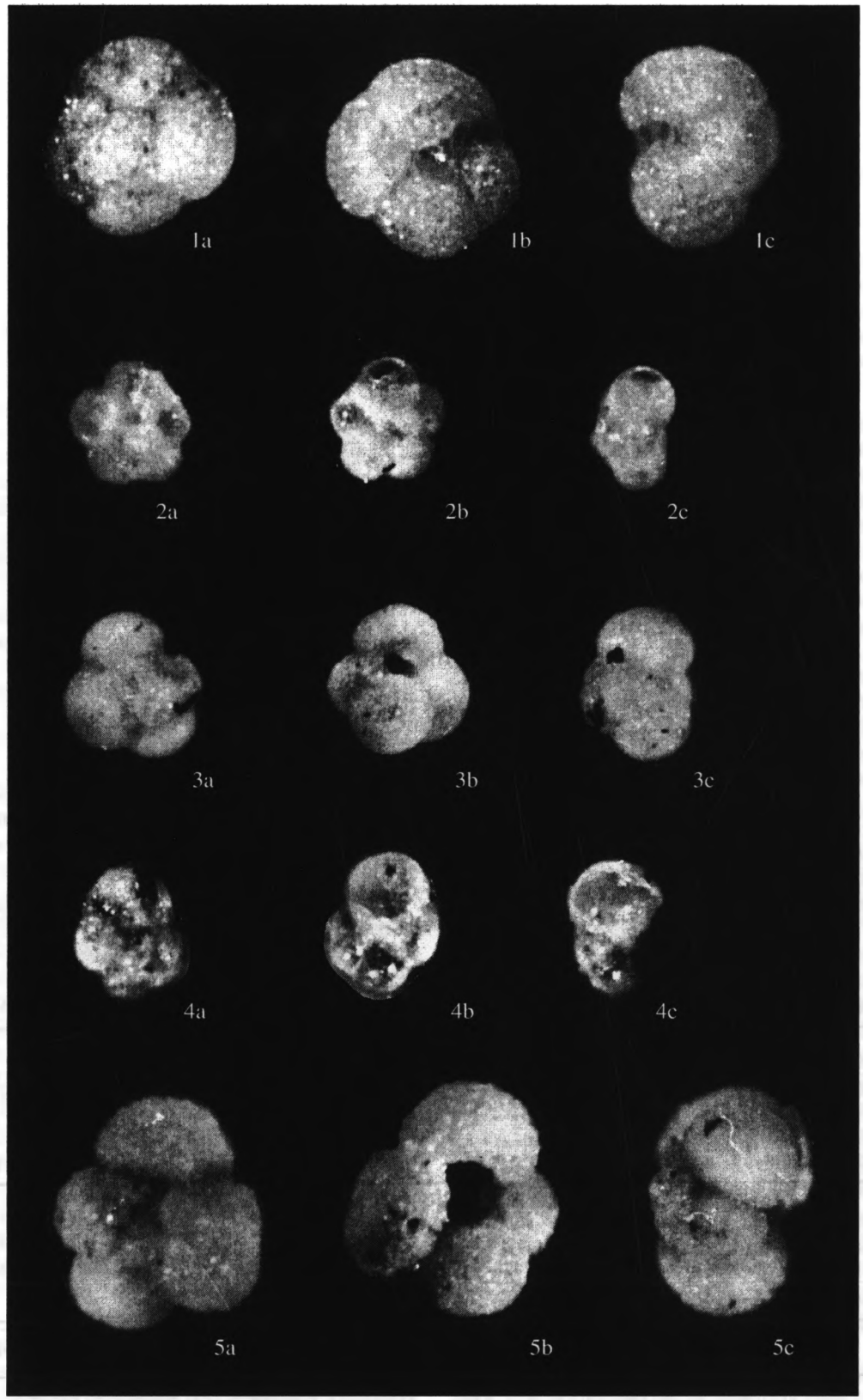
Data on Paleogene foraminifer assemblages from the Far East as a whole show that planktonic foraminifers with dominant globigerinas occur sporadically and have an impoverished taxonomic composition as compared with those from the low latitudes. The appearance of planktonic foraminifers in the Cenozoic deposits of high latitudes was mainly associated with transgressions and relative climatic warmings recorded by the oxygen-isotope data (Shackleton and Kennett, 1975; Letolle, 1979; Vergnaud-Grazzini, 1984). In the Paleogene of the Far East, these events are marked by the occurrence of calcareous plankton in the lower Paleocene, at the Paleocene-Eocene boundary, at the beginning of the early and late Ypresian, in the Lutetian, and in the lower and upper Bartonian (Serova, 1966; Serova, 1967; Serova, 1969a; 1969b; Krasheninnikov *et al.*, 1988; *Opornyi razrez* ..., 1994).

Findings of calcareous plankton in the Oligocene deposits of the Far East of Russia are very scarce. By now, planktonic foraminifers are known from the Nikol'skoe Formation of the Bering Island (Komandorskie Isles), where Serova recovered *Globorotalia pseudokugleri* Blow, later classed with *Gl. gemma* Jenkins (Krasheninnikov *et al.*, 1988), and *Gl.* sp. (Schmidt *et al.*, 1973). At the base of the Il'khatunvayam Formation, Karaginskii Island, Serova identified *Pseudohastigerina micra* (Cole) as reported by (Krasheninnikov *et al.* (1988).

According to oxygen-isotope records, the middle Oligocene (nannoplankton zones CP17-CP18) peak of warming against the background of the general decrease of surface water temperatures was manifested in the North and South Atlantic, Indian Ocean, and in the North and Central Pacific (Borshchevskii *et al.*, 1993; Corfield and Cartlidge, 1993).

According to recent data of Shcherbinina (1997), the occurrence levels of nannoplankton in the Paleogene sediments of the Kronotskii Peninsula is directly correlated with transgressive maxima of the eustatic curve of Haq *et al.* (1987). We can suggest that the occurrence of a bed rich in biogenic carbonate and calcareous plankton within the studied substantially siliceous sequence corresponds to the transgression in the second half of the early Oligocene.

It should be noted that some benthic foraminifers from this level, such as *Bolivina marginata adeloidana* Cush. et Kleinpell, *Gavelinella californiensis* (Cush. et Hobson), *Uvigerina* cf. *galloway blakeleyensis* Fulmer, *Valvulineria* aff. *willapaensis* Rau, and *Cibicidoides* aff. *elmaensis* Rau (all of them occur in the Oligocene Blakeley Formation of California studied by Fulmer, 1975), are typical of the Pacific and coexist with the fol-



Oligocene planktonic foraminifers of the Kronotskii Peninsula.

(1) *Globigerina ciperoensis angustiumbilicata* Bolli, Sample 18/19; (2) *Globigerina ciperoensis ciperoensis* Bolli, Sample 21/9; (3) *Globigerina ampliapertura* Bolli, Sample 21/9; (4) *Globorotalia opima nana* Bolli, Sample 21/9; (5) *Globigerina praebulloides praebulloides* Blow, Sample 21/9. For all images: (a) spiral side; (b) umbilical side; (c) lateral side; × 125.

lowing North Atlantic forms also characteristic of the Oligocene in Europe: *Cibicidoides lopjanicus* Mjatl., *C. tenellus* (Reuss), *Alabamina tangentialis* (Clodius), *Globocassidulina subglobosa* (Brady), *Bolivina fastigia* Cush., *Angulogerina gracilis germanica* Cush. et Edwards, *A. gracilis tenuistriata* (Reuss), *Pullenia bulloides* (d'Orb.), *Obliquina substriata* (Williamson), *O. semistriata* (Williamson), and *Fissurina soldanii* Seguenza. These forms were recorded in foraminifer assemblages of Carpathians (Myatlyuk, 1950), in septarium Clay of northern Germany (Kiesel, 1962), in the Boom Clay of Belgium (Batjes, 1958), and in the Hempsted Beds of England (Murray and Wright, 1974).

The coexisence of Atlantic and Pacific faunas and the occurrence of planktonic forms in the Oligocene of the Far East most likely resulted from a short-term warming in the middle Oligocene and from the global rise of sea level favorable for broader connections and, correspondingly, for migration of biota between the Atlantic and Pacific oceans. Assumptions agreeable to our results were made on the basis of study of the Oligocene radiolarian and diatom assemblages from the Bering Island, Komandorskie Isles (Vitukhin, 1993; Fedorchuk *et al.*, 1987).

CONCLUSIONS

(1) The detected assemblages of planktonic foraminifers and nannoplankton representative for the high latitudes were used to discriminate a unit of the lower Oligocene deposits corresponding to the *Globigerina ampliapertura* Zone and to nannoplankton zones CP17–CP18 in the Cenozoic section of the Kronotskii Peninsula.

(2) The diverse assemblage of benthic foraminifers associated with planktonic forms characteristic of Late Paleogene deposits of the Far East and adjacent Pacific regions is reliably correlated now with the chronostratigraphic scale.

(3) Within the Kronotskii Peninsula of eastern Kamchatka, we revealed an important correlation marker of the biostratigraphic and event type. Its level supposedly corresponds to the transgression and a short-term climatic warming in the middle of the Oligocene.

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Vegetation of the Western Coast of the Baidaratskaya Bay at the End of the Late Pleistocene

A. A. Andreev*, P. E. Tarasov, F. A. Romanenko**, L. D. Sulerzhitskii***, and K. I. Terekhov†, ***

**Institute of Geography, Russian Academy of Sciences, Staromonetnyi per. 29, Moscow, 109017 Russia*

***Moscow State University, Vorob'evy Gory, Moscow, 119899 Russia*

****Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia*

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Abstract—The results of palynological analysis of a section of unconsolidated deposits in the 12-m-high coastal escarpment of the Baidaratskaya Bay (68°51'N, 66°54'E), the southeastern part of the Yugorskii Peninsula, are outlined. The section includes the lower member of sandy loam rich in ice, presumably accumulated in the late glacial time. The upper peat member was accumulated between 8 and 4 ka ago. The rough boundary between peat and oxbow-lake loams with cryoturbation marks suggests a hiatus in sedimentation between 10.9 and 8.2 ka ago. The age of deposits was determined by the radiocarbon method. The results of palynological analysis reveal several phases of vegetation development in the study area and specify their time limits. We believe that the Allerod warming period was more favorable than the Middle Holocene for the development of arboreal–fritucose vegetation in the region now occupied by typical tundras. This is likely a result of the sea level rise at the end of the late Pleistocene and in the first half of the Holocene up to altitudes somewhat exceeding the recent level, when the influence of the ocean upon climate and vegetation was enhanced.

Key words: *unconsolidated deposits, terraces, peat, pollen analysis, radiocarbon dating, vegetation changes, sea-level fluctuations, late Pleistocene, Holocene.*

INTRODUCTION

The Yugorskii Peninsula and the western coast of the Baidaratskaya Bay are so far the least investigated areas of the Russian Arctic region. At the same time, their geographic position at the border between Europe and Asia, where they are open to the influence of the late Pleistocene glaciers of the Polar Urals and Pai Khoi, makes them the key areas for reconstructing the development of nature in the Arctic coast of Eurasia. New data, obtained through the interdisciplinary research of unconsolidated deposits of the organic–mineral Baidara section, greatly contribute to a better understanding of the evolution of vegetation and natural processes in the region during the period from the late Pleistocene up to the 4 ka ago.

Natural Environment in the Study Area

Our investigations were carried out in the southwestern coast of the Baidaratskaya Bay (Fig. 1). It is a shallow bay of the Kara Sea extending deep into the land between the Yugorskii and Yamal peninsulas. The surface of the coastal plain is flat or gently inclined, with abundant swamps and lakes. The Quaternary sediments are predominant sands with interbeds of loam, plant detritus, and peat. Northern spurs of the Polar Urals (altitude 500–1300 m) and Pai Khoi Ridge (467 m) are situated at a distance of about 50–80 km southwest and west of the section, and diminish the influence of the Atlantic air mass. As reported by the Ust Kara meteostation situated 100 km farther northwest, the July mean temperature is 7°C, the mean temperature in January is –22°C, and the frost-free period is 67 days. The precipitation in July is about 50 mm, the mean annual precipitation is 425 mm (*Spravochnik...*, 1967, 1969). As the 50 m isobath is detected at a distance of 150–200 km northward, near the Bay mouth, it may be suggested that the continental climatic parameters around the section were intensified at the very end of the late Pleistocene during the low sea-level stand. Low summer temperatures and relatively high humidity are responsible for a poor recent vegetation that is represented by moss, herbaceous, and fruticose tundras. Adjacent areas, occupied by the pretundra zone of thin forests of Siberian spruce and arborescent birch, are situated at a distance of 200 km south-southeast of the section, whereas fruticose alders are typical of the Pai Khoi and Polar Urals.

Samples for the palynological, botanical, and radiocarbon analyses were collected from the Baidara sec-

† Deceased.

tion (68°51′N, 66°54′E) in September–October, 1993. The upper 10-m-thick organic–mineral part of the coastal escarpment of the Baidaratskaya Bay (12 m high in total) is under the influence of intense wave abrasion. The lower part of the escarpment is covered by talus and unsuitable for sampling. The section is located near the border of an ancient lacustrine depression. In the northern part, the depression surface was cut off by wave abrasion, and its eastern part was affected by lateral erosion of a small river, flowing from south to north. Another river (Ngoyuyakha), whose source is in the Polar Urals, flows into the Baidaratskaya Bay, 3 km west of the section. In the uppermost part, the surface of the coastal escarpment is dissected by fractures (about 0.8 m wide and 0.6 m deep), formed above thawing ice wedges.

RESULTS AND THEIR INTERPRETATION

We present the obtained results by describing in parallel the lithologic–geochemical characteristics of the deposits and the composition of spore–pollen spectra. We believe that this is convenient for a better understanding of specific features of sedimentation and for a explanation of the origin of pollen and spores incorporated in sediments.

Late Glaciation Time

The lower part of the section at the depth of 6.2–10.3 m is mostly represented by gray and yellow–gray fine- to medium-grained sands with interbeds of coarse sand, gravel, and bluish loam (Fig. 2). Subhorizontal lamination (interbeds are about 1 cm thick) is characteristic of these deposits. The origin of gravel interbeds is of particular interest. The pre–Quaternary conglomerates at the base of the coastal cliffs may be regarded as their source. Changes in the granulometric composition of sediments throughout the section; alternation of sandy–gravel and loam beds; and a relatively high content of SO_4^{2-}, HCO_3^-, and Ca^{2+} ions in the extracted fluid suggest the continental genesis of the deposits. Konyakhin *et al.* (1991), who investigated the deposits in the 8–15-m-high terrace between the Baidara section and the Ngoyuyakha River mouth, arrived at the same conclusion. They consider the low content of easily soluble salts in the water extract (from 0.045% in sands to as much as 0.345% in clays) as evidence against the marine origin of grounds. Analyzing the diatom composition in samples collected from the lower part of the section, Z.V. Aleshinskaya (personal communication) suggested the alluvial origin of sediments.

Spore–pollen spectra in deposits at the depth of 6.2–10.3 m (Fig. 2) contain the rather abundant pollen of *Betula* sect. *Albae*, *Abies*, *Picea*, *Pinus* s/g *Haploxylon*, *Pinus* s/g *Diploxylon*, *Tilia*, *Corylus*, and *Juglans*. Spectra of a similar composition were detected in deposits of the Mikulino (Kazantsevo) time of the Pai

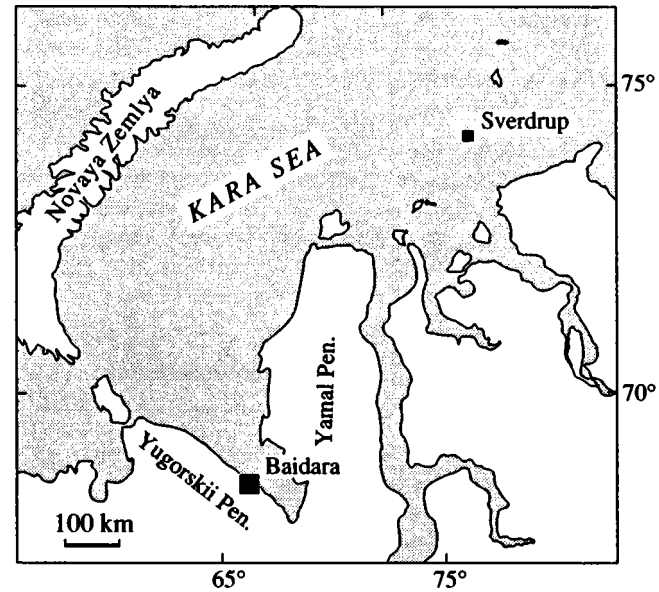

Fig. 1. Schematic map of the region with indicated localities of the Baidara and Sverdrup sections.

Khoi Ridge (Duryagina and Konovalenko, 1993); this may imply that a considerable part of spores and pollen was redeposited from older Pleistocene deposits. Some of pollen grains (*Juglans*, for example) may be pre-Quaternary in age. Contamination of glacial deposits with pollen of thermophylic plants redeposited from interglacial sediments was repeatedly mentioned in the literature, in particular, this is characteristic of the Russian plain (Chebotareva and Makarycheva, 1974). The criterion of identifying the redeposited pollen is its ecological incompatibility with pollen of local plants, such as dwarf birch, sedge, wormwood, grasses, heath, and of some others, regularly present in spectra in small amounts.

Opportunities for dating the above-described bed are relatively restricted. Konyakhin *et al.* (1991) dated the age of the 8–15 m terrace as late Pleistocene (Sartanian). The buried stratal ice at the base of the terrace is exposed along the shore for a considerable distance. The maximum thickness of the ice is 5 m, and its visual extension is 70–80 m. No ice sheets have been recovered at the Baidara section. This suggests a younger, late glacial age of the studied deposits.

The glaciation center that existed in the Pai Khoi Ridge and Polar Urals 18–20 ka ago and was the source area for glaciers spread up to the southwestern Yamal was suggested by many authors (*Razvitie landshaftov* ..., 1993). However, problems concerning the time of deglaciation and complete disappearance of the ice cover in the region still remain to be clarified.

In late glacial time (16–12 ka ago), the southeastern part of the Yugorskii Peninsula represented, most likely, an open landscape similar to recent arctic deserts. Under conditions of a severe climate, the very sparse

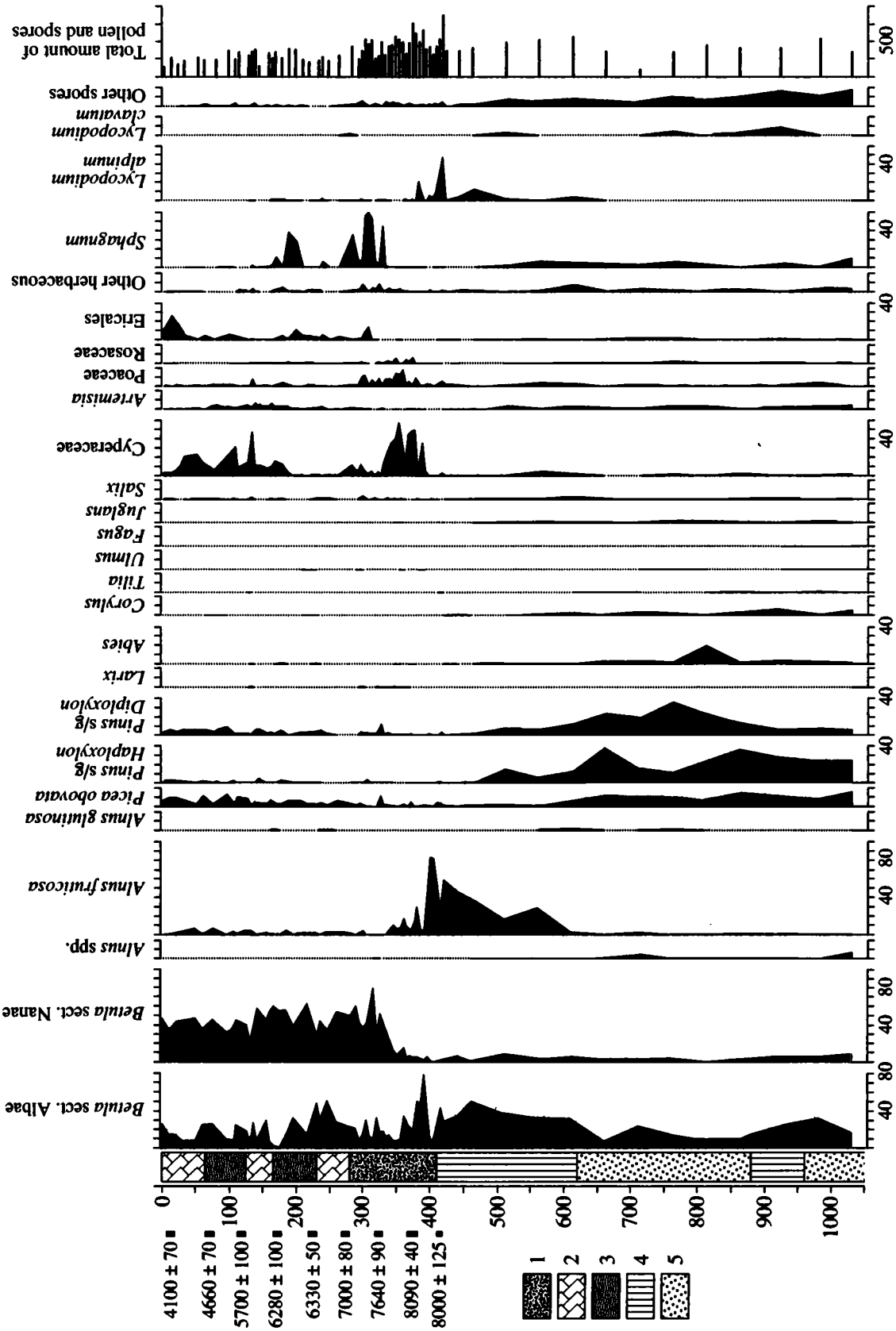

Fig. 2. Spore–pollen diagram for the Baidara section (percentages of pollen and spores are calculated relative to the total taxa amount in the spectrum: (1) sphagnum peat, (2) sedge–sphagnum peat, (3) sedge peat, (4) silty–pelitic sediments (loams and sandy loams), (5) sands with gravel interlayers.

vegetation did not produce an amount of pollen sufficient enough to be dominant in the spectra. Thawing of mountain glaciers intensified the river run-off, and the low sea level favored the erosion and redeposition of more ancient (pre-Quaternary including) sediments, as is evident from gravel interbeds and thermophylic pollen present in the section.

The Alleród Time

The silty-pelitic sediments occur higher in the section at the depth of 4.2–6.2 m. Their intense fragmentation and cryoturbation (wavy–lenticular cryostructure) suggest that they were accumulated under sufficiently severe climatic conditions.

The abundant pollen of *Alnus fruticosa* Rupr., arborescent birch, and spores of *Lycopodium alpinum* L. are characteristic of this interval. Simultaneously, the share of redeposited arboreal pollen and spores notably declines here (Fig. 2).

The upper horizon of the unit extending along a considerable part of the shore is represented by thin lenses of loam and sandy loam enriched in remains of alder stems and branches up to 4–6 cm in diameter. Their fragments are usually rounded in the course of fluvial reworking. Taking into account the abundance of alder pollen in deposits, we can hardly suggest redeposition of wood remains from older sediments. The radiocarbon analysis of a branch (about 6 cm in diameter), collected in 1991, yielded the age date of 10900 ± 120 years (MGU-1362). This implies that the silty–pelitic sedimentary unit can be attributed to the Allerod. The rough boundary with the overlying peat bed, marked by cryoturbation marks, suggests a hiatus in sedimentation in the past oxbow lake since the Younger Dryas. The hiatus may correspond to the time of local rearrangement in the hydrographic network and lake shoaling under conditions of climatic cooling and growing aridization combined with the low erosion base.

Our interpretation of spore–pollen spectra suggests that the thin birch forests of the forest–tundra type with a considerable alder admixture surrounded the studied locality in the Allerod time (about 12–11 ka ago). Dwarf birches (*Betula nana* L.) were also components of the vegetation cover. As was mentioned above, the present zone of arboreal vegetation is located farther to the south.

Holocene

The silty–pelitic member is overlain by a thick (0.0–4.2 m) peat bed. In coastal cliffs of the southwestern part of the Baidaratskaya Bay, the peat bed is discontinuous and consists of lenses extending for several dozens of meters. Radiocarbon dates for samples collected from the lower part of the peat are as follows: 80000 ± 125 years at the depth of 4.1–4.15 m (WAT-2924); and 8210 ± 110 (GIN-7862a), 8120 ± 90 (GIN-7862b), and 8090 ± 40 years (GIN-7862c) at the depth of 3.7–3.85 m. These dates suggest that the peat accumulation started about 8000–8200 years ago. Therefore, the recorded stratigraphic hiatus falls within the interval of 10900—8200 years ago.

In general, pollen spectra at a depth of 3.4–4.2 m resemble those of the Allerod time. The share of Cyperaceae, Poaceae, and Rosaceae is lower in this interval, where we detected simultaneously increased abundances of *Alnus fruticosa* pollen and *Lycopodium alpinum* spores.

The radiocarbon date of 7640 ± 90 years (GIN-7869) for the peat sample from the depth of 3.4–4.2 m corresponds to the end of the Boreal–early Atlantic period of the Holocene (about 8200–7700 years ago). The complete disappearance of alder pollen from the spectra is recorded at about 7700 years ago. According to palynological data, the vegetation of that time in the area under study was steadily dominated by birch forests of the forest–tundra type, but as compared with the Allerod time, the alder population considerably declined in them. This can be explained, apart from climatic reasons, by progradation of bogy landscapes in the area that was responsible for the accumulation of the lowland sedge peat. By the Atlantic time, the sea level rose up to the recent stand (Popov *et al.*, 1988) and, as a result, the erosion base and ground water level rose also. Moreover, the activation of peat-forming processes and an increase in the peat accumulation rates in the Atlantic period are characteristic of the entire northern Eurasia, which may be a result of the optimal ratio between the heat and humidity (Khotinskii, 1977).

From the depth of 3.40 m up to the surface of the section, the spectra are dominated by dwarf birch pollen and indicate the prevalence of fritucose coenoses in the peat bog landscape. The alder pollen is persistently present in spectra in small amounts, showing the insignificant role of these plants in the vegetation cover after 7700 years ago. Nevertheless, macroremains of alder discovered at a depth of 2.2–2.3 m from the peat bed 6330 ± 50 years old (GIN-7868) confirm its presence at the surface of the peat bog.

The content of arboreal birch pollen in the spore–pollen spectra periodically varies, declining up the section. The minimum content was recorded in the depth interval of 1.65–1.75 m marked by the radiocarbon date of 6280 ± 100 years (GIN-7860). We believe that variations in the content of arboreal birch taxa reflect the degree of their participation in the vegetation cover of the region that was mainly controlled by the temperature regime. Thus, the minimum content recorded 6300 years ago agrees well with the appreciable cooling that was typical of other regions of northern Eurasia, according to palynological evidence (Andreev *et al.*, 1989; Andreev and Klimanov, 1991). The abundance peaks of *Betula* sect. *Albae* pollen are recorded in peat layers that we refer to the last third of the Atlantic period of the Holocene (4500–6000 years ago). Radiocarbon dates of 5700 ± 100 years for the depth interval

of 1.15–1.25 m (GIN-7867) and of 4660 ± 70 years for the depth interval of 0.65–0.75 m (GIN-7868) allow us to correlate the distinguished peaks with warming phases which occurred 6000, 5500, 5200, and 4600 years ago. A notable decrease in the abundance of *Betula* sect. *Albae* pollen is registered in the depth interval of 0.25–0.5 m. We correlate this level with a period of cooling that occurred about 4500 years ago (Khotinskii, 1977). Another warming marked by an increase in abundance of arboreal birch pollen took place 4000 years ago.

Changes in composition of herbaceous pollen and spores reflect the variability of coenoses at the peat bog surface. A high content of sphagnum spores in a depth interval of 1.75–3.40 m, associated with sedge and Ericaceae pollen, suggest an important contribution of sphagnum moss to the peat bed accumulation about 6300–7700 years ago. The final period of the Atlantic time is characterized by the nearly complete disappearance of sphagnum spores from the spectra predominantly enriched in sedge pollen. Sedge coenoses became widespread again at the peat bog surface 6300 years ago. The peak content of sedge pollen in the upper 20-cm-thick peat layer (4140 ± 70 years old in the depth interval of 0.15–0.2 m, WAT-2895) indicates another change in coenoses at the peat bog surface. It might have been associated with a transformation of the hydrological regime responsible for the bog desiccation and break in the peat accumulation about 4000 years ago.

The spruce pollen that is persistently present in spectra from peat layers increases in abundance during the last third of the Atlantic period. This indicates the northerly migration of the spruce distribution areas at that time.

Apparently, the vegetation cover around the peat bog in the middle Holocene consisted of thin birch forests and fruticose tundra areas alternating with open boggy landscapes.

A series of radiocarbon dates of 7000 ± 80 (GIN-7864), 6390 ± 50 (GIN-7865), 6280 ± 100 (GIN-7860), 5700 ± 100 (GIN-7867), 4660 ± 70 (GIN-7868), and 4140 ± 70 years (WAT-2895) was obtained for peat samples from different depth intervals. These dates indicate that peat accumulation continued during the entire Atlantic period with a rate of 1 mm/year and ceased about 4000 years ago.

Popov *et al.* (1988) distinguished a transgressive phase that occurred 6000–4500 years ago, when the sea level was presumably 2–3 m higher relative to the recent stand. The regressive phase took place 4500–2000 years ago. The conditions of a high sea-level stand and climatic warming of the Atlantic period were responsible for the expansion of boggy landscapes. The majority of bogs became dry as a result of sub-Boreal regression.

CONCLUSIONS

The palynological data obtained from the Baidara Section clearly indicate changes in vegetation of the southeastern part of the Yugorskii Peninsula at the end of the late Pleistocene and in the middle Holocene. Open landscapes with sparse vegetation (similar to modern arctic deserts) typical of the late glacial time were replaced during the Allerod by light birch forests of the forest–tundra type with a considerable admixture of alder and dwarf birch.

A hiatus in sedimentation in the interval between 10900 and 8200 years ago made it impossible to define the composition of vegetation at that time.

At the Boreal–Atlantic boundary (the Holocene), arboreal and fruticose birch taxa still prevailed in the vegetation, while the amount of alder notably declined. This presumably was associated with the increased development of bogs.

During the main part of the Atlantic and in the early sub-Boreal periods (7700–4000 years ago), vegetation was represented by birch tundra and light forests, alternating with open boggy areas mostly occupied by sedge and sphagnum assemblages.

Peat accumulation ceased about 4000 years ago, when the sub-Boreal regression of the Kara Sea started. At present, the studied locality is surrounded by typical tundra landscapes.

We think that the recognized changes in vegetation were controlled not only by the general evolution of the climate, but also by the position of sea level. The climatic warming of the Allerod time, when the sea level was rather low and the vast shelf areas of the Kara Sea represented the dry land, displaced the boundary of the arboreal–fruticose vegetation much further to the north than the warm Atlantic optimum, when the sea level was higher than now. Earlier we reported about similar events in the Sverdrup Island (the Kara Sea), which is now occupied by the sparse vegetation of the arctic desert. Palynological data obtained from the buried peat bog suggest that the island was covered with herbaceous and fruticose tundras 12–9.5 ka ago (Tarasov *et al.*, 1995).

As is evident from palynological data, the short-term climatic fluctuations (primarily, fluctuations of summer temperature under the sea cooling influence) were also characteristic of the middle Holocene. We distinguished and dated by the radiocarbon method the warming phases, which occurred about 6000, 5500, 5200, and 4000 years ago and were separated by cooling episodes.

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BRIEF COMMUNICATIONS

Fossil Assemblages from the Devonian Shallow-Water Deposits of Eurasia (European Russia and Mongolia)

V. A. Aristov

Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Shallow-water deposits of Paleozoic epicontinental seas on ancient platforms along with those of shelf basins in marginal parts of paleocontinents are characterized by certain peculiar features, which distinguish them from the deposits of relatively deep water basins. Distinctive characteristic of the former are the rhythmic alternation of terrigenous and carbonate rock members in the section corresponding to the alternation of transgressive and regressive cycles of sedimentation and also the presence of hiatuses at the cycle boundaries. The taxonomic composition of biotic assemblages also changes in accord with changes in the lithology of the sequences, because the habitats of marine organisms are closely interrelated with the sedimentation settings.

The analysis of biotic changes in Devonian shallow-water deposits was performed in the Russian platform and Variscides of southern Mongolia. The bed-by-bed distribution of macroremains observable in the course of visual investigation of the section and microfossils from insoluble residues of rocks subjected to acidic treatment were considered during the analysis. The presence or absence of fossil assemblages was taken into account without their division into taxons different in rank and without their quantitative analysis. The only exception was made for conodonts, whose distribution was analyzed at the family and genus levels.

Studies of this kind for different periods of the Phanerozoic are well known from the world literature. In particular, investigation of the Devonian was performed by R.F. Gekker and V.S. Sorokin. The detailed study of the Frasnian shallow-water deposits in the southwestern Russian platform (Main Devonian Field) allowed Sorokin to establish regular changes in bioecological assemblages and facial types of deposits both upward the sequence (from one rhythm to another) and in the lateral directions. In addition, he defined the stratigraphic significance of separate species and groups of species (Sorokin, 1978).

Materials presented in the paper are insufficient to analyze the lateral distribution of biotic assemblages, because they concern only separate sections of two regions, where studies of this kind have not been performed. The results obtained do not contradict conclusions by Sorokin, but the analysis of conodonts repesenting the microfossil group indicative of ecological–facial environments supplements previous studies.

In shallow-water basins, the posthumous displacement of organic remains from the dwelling site to the place of burial is insignificant. Therefore, fossil biotic assemblages reflect, to a significant extent, a structure of living assemblages in the respective part of paleobasin.

MIDDLE–UPPER DEVONIAN BIOTIC ASSEMBLAGES OF THE RUSSIAN PLATFORM

Biotic assemblages were analyzed in two sections recovered by Boreholes 178 (Srednii Karachan Village) located and 76 (Bol'shie Alabukhi Village) at the northeastern slope of the Voronezh anteclise (the Central Devonian Field), 200 and 215 km east–southeast of Voronezh, respectively, and supplementing each other. The composite section 410 m thick spans the time interval from the late Eifelian (the Klintsovo Horizon) to late Famennian (the Ozerskii Horizon). Substantiation of individual subdivisions was discussed earlier (Aristov, 1988). Local stratigraphic units for the Central Devonian Field of the Russian platform are shown in Figure 1 (the left-hand part). At the base of the section, the Klintsovo terrigenous deposits overlie with a structural unconformity the metamorphosed rocks of the Proterozoic basement. Higher in the section, there is an alternation of carbonate and terrigenous rock members. Six cycles are distinguished in the section: one in the Eifelian, two in both the Givetian and Frasnian, and one in the Famennian. Three hiatuses recorded at the cycle boundaries are identified as pre-Givetian (the Chernoyarsk Horizon is partly washed out), pre-Frasnian (the sandstone bed with ferruginous oolites occurs above the hiatus), and pre-Voronezh (the Petino Horizon is missing from the section). A cryptic hiatus probably occurs at the base of the Zadonsk Horizon. A stage of

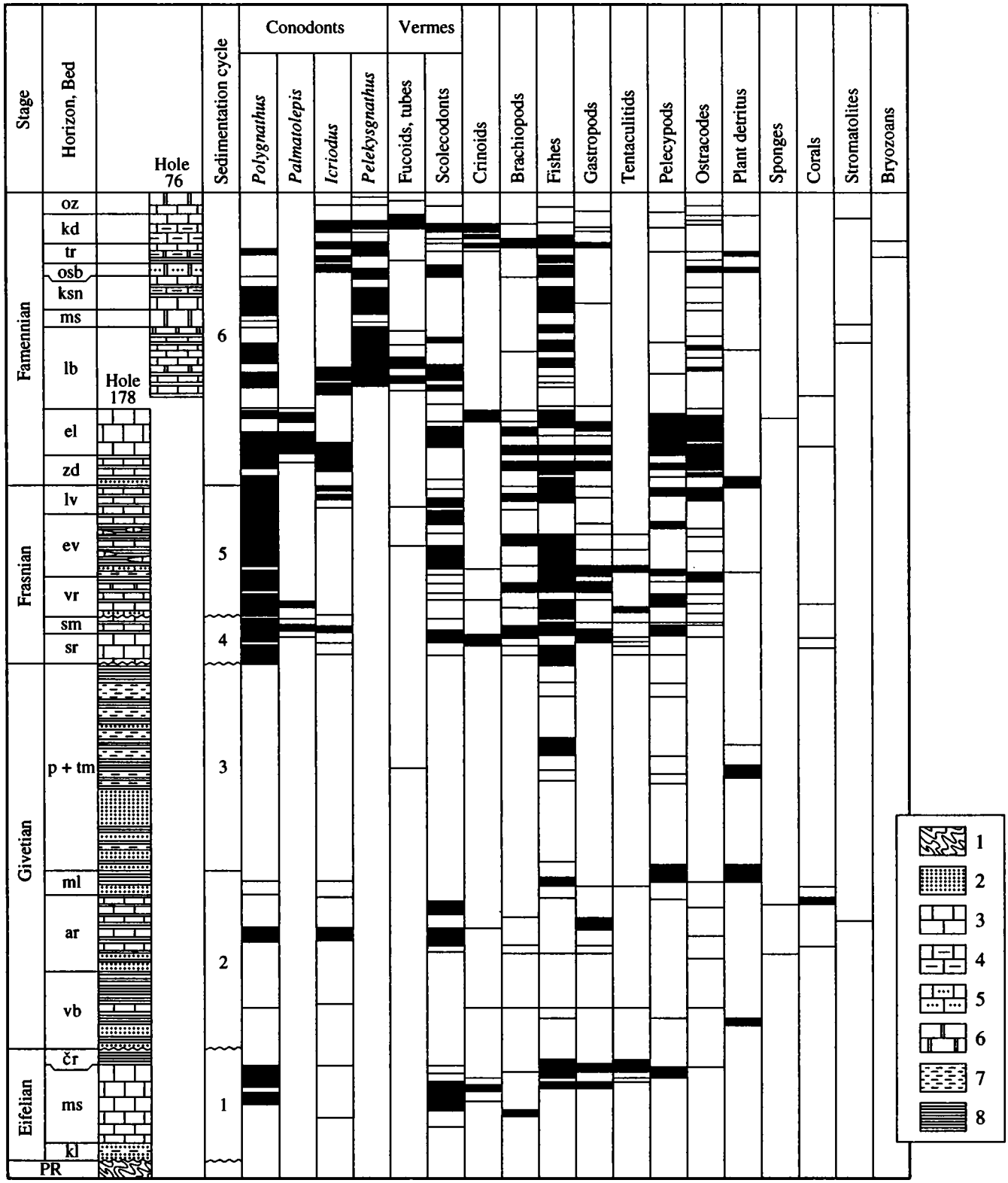

Fig. 1. Middle–Upper Devonian biotic assemblages from the Central Devonian Field (Russian platform): (1) Proterozoic basement rocks; (2) sandstone; (3–5) crystalline, organogenic, or detrital limestone; (4) clayey limestone; (5) sandy limestone; (6) dolomite; (7) siltstone; (8) shale.

Horizons: (kl) Klintsovo, (ms) Mosol, (r) Chernoyarsk, (vb) Vorob'evo, (ar) Ardat, (ml) Mulla, (p) Pasha, (tm) Timan, (sr) Sargaevo, (sm) Semiluki, (vr) Voronezh, (ev) Evlanovo, (lv) Livny, (zd) Zadonsk, (el) Elets; (lb) Lebedyan', (oz) Ozerskii; beds: (mc) Mtsensk, (ksn) Kiselevo–Nikol'skoe, (osb) Orel–Saburovo, (tr) Turgenevo, (kd) Kudeyar.

the transgression onset, the transgressive stage proper, and regressive stage are distinguished in every cycle.

The quantitative analysis of organic remains was carried out at 195 levels with an average step of observations approaching 2.1 m. However, the frequency of observation points in relation to the lithologic composition of rocks was irregular: relatively rare in the Eifelian and Givetian and more frequent in the Frasnian and Famennian. Fourteen groups of organic remains are present throughout the section: conodonts, worms (scolecodonts, fucoids, and tubes), crinoids, brachiopods, fish (teeth, scales, bone fragments), gastropods, tentaculitids, pelecypods, ostracodes, corals, bryozoans, stromatolites, and plant remains. Distribution of these fossils is shown in Figure 1, where thin lines mark the occurrence of representatives of any group at a single level, and black intervals indicate several successive finds in the section.

Conodonts are spread almost over the entire studied section: at separate levels in the Eifelan and Givetian and virtually in every Upper Devonian bed of the section. The exception is the upper half of the Givetian barren of conodonts. The four most characteristic genera taken into consideration among conodonts were *Polygnathus*, *Palmatolepis* (both representing polygnathids), *Icriodus*, and *Pelekysgnathus* (icriodids). Representatives of the most widespread *Polygnathus* genus occur in both the transgressive and regressive stages of cycles. Thus, the genus is not an indicator of particular environments. The difference in the distribution of its representatives can be detected only at the species level or at the level of group of species (Aristov, 1988).

Another polygnathid genus *Palmatolepis* characteristic of the Upper Devonian occurs only at the levels corresponding to the transgressive stages (cycles 4–6).

Icriodids are also present in both the transgressive and regressive facies. However, they are more diverse and abundant in deposits of regressive stages. Representatives of the *Pelekysgnathus* typical of the entire Devonian occur in the Central Devonian Field only in the Upper Famennian portion of the section.

Other fossil groups are distributed in the section in a different manner. Nektonic (fish) and infaunal (worms) species, as well as tentaculitids, pelecypods, and ostracods are only slightly dependent on the environmental factor and occur through the entire section more or less regularly. Crinoids, brachiopods, and corals are mainly recorded in deposits corresponding to transgressive periods. Gastropods and plant remains, in contrast, are more characteristic of regressive sedimentation stages. Among others rarely occurring groups of organic remains, there are sponges (their spicules are recorded at three levels in transgressive facies of cycles 2 and 6), stromatolites (4 levels in regressive facies of cycles 2 and 6), and bryozoans (2 levels in the regressive deposits of cycle 6).

The most impoverished faunal assemblages are found in the upper Givetian portion of the section (Pasha and Timan horizons). This time was marked by the preponderance of subcontinental sedimentation settings. Corresponding sediments incorporate only fish remains, pelecypods, and plant detritus.

LOWER–MIDDLE DEVONIAN BIOTIC ASSEMBLAGES OF MONGOLIA

The second object of investigation was the Tsakhirin Khuduk section located on the southern slope of the Gobi Altai in southern Mongolia, 10 km southeast of the Shine Dzhinst Somon near the Tsakhirin Khuduk well. In the structural respect, the region is referred to the Dhinset subzone of the Gobi–Altai–Sukhebator zone of South Mongolian Variscides (Ruzhentsev and Pospelov, 1992). The section spans the Lower–Middle Devonian part of the thick Paleozoic vocanogenic–sedimentary sequence.

The Tsakhirin Khuduk section is composed of terrigenous and carbonate rocks with rare volcanomictic interbeds (tuffaceous sandstones and tuffites of acid and intermediate composition). By lithology, these rocks represent the relatively shallow shelf deposits. The section is subdivided into the Tsakhirin, Chulun, and Tsakhangal horizons and Givetian beds (Alekseeva, 1990); all these units are dated by benthic fauna.

The Tsakhirin Horizon of presumably Pragian age is represented by a 500-m-thick member of limy conglomerates, sandstones, and siltstones unconformably resting upon different Silurian rocks. Single finds of redeposited Silurian conodonts are recorded in limestone pebbles. The overlying Tsakhangal Horizon was traditionally considered to be the Eifelian in age. The results of conodont studies, however, allow the lower part of the section to be referred to the upper Emsian.

In the stratigraphic respect, the section is incomplete, and the upper Eifelian deposits are missing from it because of tectonic distortion.

The section reveals, similarly to the section of the Russian platform, a rhythmic alternation of transgressive–regressive cycles. Three major units recognized correspond to the Pragian–lower Emsian, upper Emsian–Eifelian, and Givetian cycles.

Fourteen groups of organic remains were taken into consideration when analyzing biotic assemblages: conodonts, tabulates, rugoses, gastropods, stromatoporates, ostracodes, crinoids, bryozoans, fish, tentaculittids, nautiloids, sponges, and trilobites. Occurrence levels of fossils representing all these groups are shown in Figure 2. Their occurrence frequency in the section is lower compared with the Central Devonian Field (117 levels per 1000 m of the section spanning the Emsian–Givetian interval). More frequent are conodonts, brachiopods, crinoids, bryozoans, and ostracods. All the groups are distributed throughout the section unevenly, in separate intervals. The latter are generally nar-

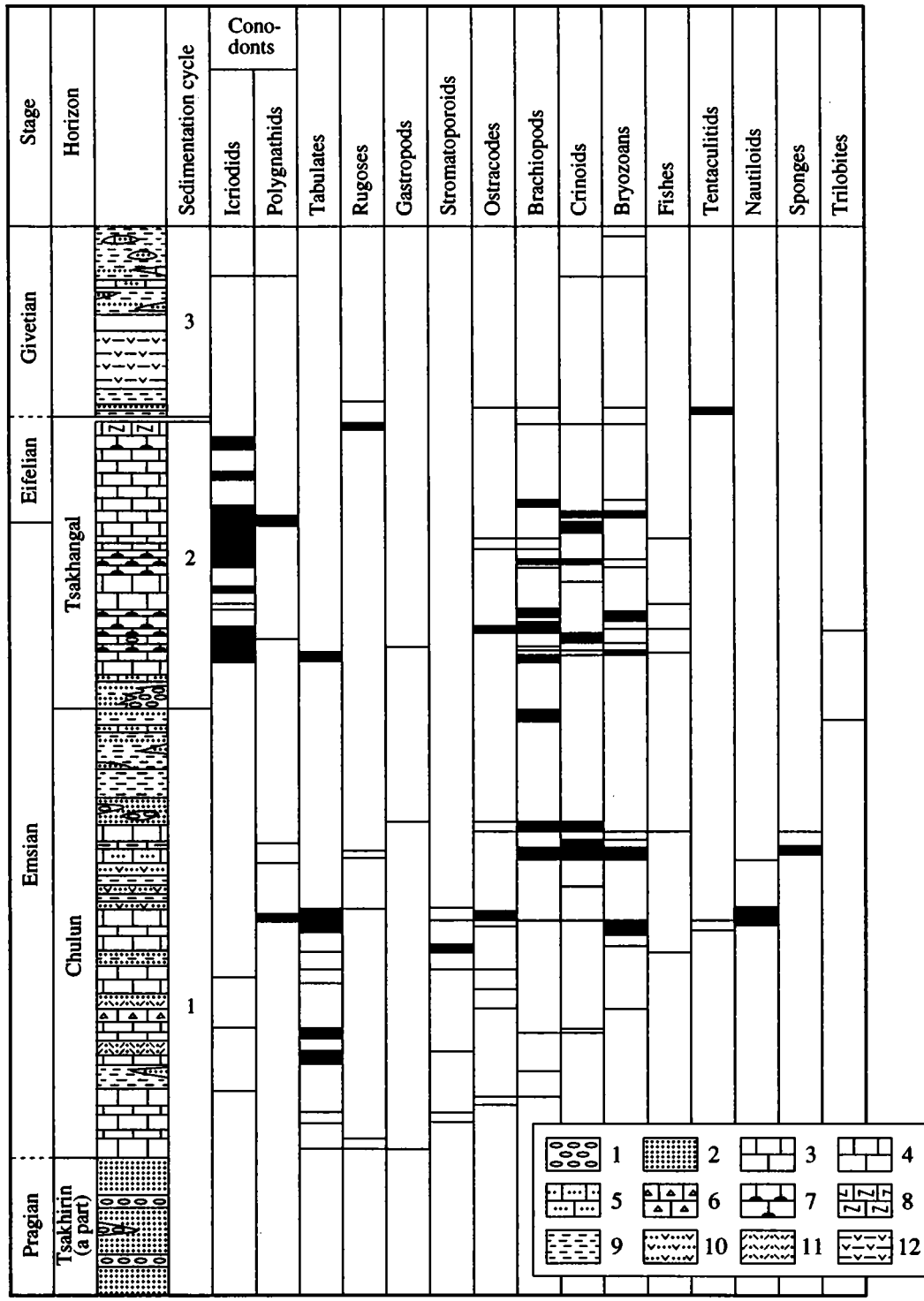

Fig. 2. Middle–Upper Devonian biotic assemblages from the Tsakhirin Khuduk section (southern Mongolia).
(1) Conglomerate; (2) sandstone; (3) organogenic, or detrital limestone; (4) clayey limestone; (5) sandy limestone; (6) brecciated limestone; (7) limestone with siliceous concretions; (8) siliceous limestone; (9) siltstone, argillite; (10) basic and intermediate tuffs; (11) acid tuff; (12) ashy tuff.

row and isolated, but sometimes the occurrence levels can be grouped into dense series spanning the more or less wide intervals of the section. Such distribution patterns reflect the unstable habitat conditions in palobasin through time.

Conodonts are mainly represented by icriodids (they are particularly abundant in the Tsakhangal Horizon) and represent the background forms of the conodont assemblages. Against this background, rare lev-

els with single polygnathids are distinguished in the transgressive facies of cycles. These very levels are marked by rare finds of nautiloids and sponges, which can be considered as indicators of transgressive sedimentation.

Tabulates and stromatoporoids are characteristic of the lower Emsian. Among rarely occurring groups, gastropods, tentaculites, fish, and trilobites should be mentioned in addition to the above-mentioned nautiloids and sponges.

No distinct correlation with transgressive or regressive sedimentation phases is observed in the distribution of brachiopods, rugoses, and, to a lesser extent, of crinoids, bryozoans, and ostracods. I noted only the more frequent occurrence of representatives of these groups in transgressive facies.

CONCLUSION

As follows from the aforesaid, the distribution of fossil remains of different biotic groups throughout the section of the Devonian shallow-water deposits is not regular, but they occur sporadically at isolated levels against the background of alternation of transgressive–regressive sedimentation cycles. Such a distribution of fossils reflects unstable conditions in the paleobasin through time.

Transgressive phases of sedimentation cycles (particularly transgression peaks) can be recognized using, in addition to lithologic markers, some biotic events, for instance, the appearance of polygnathids in the section (Emsian–Eifelian of Mongolia), or the occurrence of their particular genera and species groups (Middle–Upper Devonian of the Russian platform) in association with sponge spicules, nautiloids, and some brachiopods. Stromatolites, gastropods, and plant remains are indicative of regressive phases. Nektonic forms (fish), infaunal species, and, to a lesser extent, pelecypods and ostracods have been insignificantly affected by environmental changes and occur more or less regularly throughout the section.

It should also be noted that organic remains are more abundant and diverse in transgressive carbonate deposits than in predominantly terrigenous sediments of regressive stages. These distribution patterns result, however, not only from ecological peculiarities, but from different sedimentation rates as well. The sedimentation rate of terrigenous sediments (rudaceous in particular) was higher, as compared with carbonate deposits, and, correspondingly, the relative amount of organic remains is lower in the former.

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On Isotopic Dating of Sedimentary Rocks

A. G. Rublev

All-Russia Institute of Geology, Srednii pr. 74, St. Petersburg, 199026 Russia

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Abstract—Different approaches to the Rb–Sr dating of clayey rocks are discussed, and arguments against using the whole-rock samples for this purpose are present. Much more effective and methodologically correct is the dating of fine illite subfractions. Some disadvantages of this method and its poorly elaborated aspects are also indicated. A more extensive isotopic–geochronological investigation of volcanogenic and volcano-sedimentary sequences is suggested to be a promising alternative for solving some basic problems in stratigraphy.

Key words: *Rb–Sr method, clayey rocks, illite.*

The journal *Stratigraphy and Geological Correlation* has recently published a dispute on one of the most difficult and acute problems of modern geochronology concerning the dating of sedimentary rocks (Vinogradov, 1997; Gorokhov and Mel'nikov, 1997). The dating of these rocks is particularly important, because they are exactly the objects that can be correctly coordinated with the chronostratigraphic scale, which makes them potentially the most opportune rocks for establishing an adequate scale of geological time.

It is obvious that the most reliable chronostratigraphic dates can be obtained from the marine sedimentary rocks with paleontological characteristics, and yet it is the isotopic dating of these rocks that appeared to be such a difficult task. The chemogenic sediments lack the mineral–geochronometers traditionally used in isotopic geochronology and cannot be used for this purpose, whereas the clastic sediments include components that retain the isotopic memory of their provenance. The situation is aggravated by the postdepositional transformations of sediments and by the relatively low resistance of their mineral geochronometers against the epigenetic processes.

In this respect, the close attention of the Editorial Board of the journal *Stratigraphy and Geological Correlation* to the different aspects of the problem is salutary. I extend my thanks to the members of the Board for granting me an opportunity to get acquainted with the papers on this topic prior to their publication and to take part in the discussion.

The crucial problem lies in the different views expressed on the use of the Rb–Sr method for the geochronological study of rocks. Thus, V.I. Vinogradov suggests that the whole-rock samples are appropriate for the analysis (Vinogradov *et al.*, 1994; Vinogradov, 1997). If, as a result, a statistically meaningful straight line can be drawn through the data points on an isochron plot, then it is considered to be the isochron. The isochron inclination angle is used to calculate the age attributable to a certain stage of epigenetic transformations, because "the age of sedimentation, or early diagenesis of sediments, apparently, cannot be determined at all." The alternative explanation of the obtained straight line as the mixing line, according to Vinogradov, seems to be possible only in the case, when the mixing components "are found or imagined" (!) (Vinogradov, 1997).

Gorokhov and Mel'nikov (1997), on the other hand, insist on a complex isotopic and mineralogical study of illites from the clayey rocks and on the differentiation of the noncogenetic fractions of these minerals subsequently used for dating. Moreover, the interpretation of the obtained results and the elucidation of their geologic significance can be done by using all the available geological and stratigraphical information.

On the whole, I share the viewpoint of Gorokhov and Mel'nikov, but still consider it necessary to state my opinion on some aspects not mentioned in the previous discussion.

Dating of Whole-Rock Samples of Sediments

Gorokhov and Mel'nikov suggest the isochron model for whole-rock samples is valid under four conditions: (1) all samples are coeval; (2) their intitial Sr isotope composition $(^{87}Sr/^{86}Sr)_0$ is identical; (3) the absence of Rb and Sr migration within the sample's volume (an isolated system); (4) all data points can be approximated by one straight line. It should be noted, however, that item 4 is not a condition, but a consequence of the isochron model conditioned by the first three points. In other words, Vinogradov considers it necessary and sufficient to have not the immanent properties of the isochron model, as his opponents believe, but the consequence resulting from the model. Thus, one of the basic principles of scientific research is vio-

lated: the facts that form the basis of a hypothesis cannot serve as the proof of its authenticity. In reality, however, the straight line corresponding to data points of whole-rock samples only suggests their correspondence to the isochron model which is yet to be proved. This requirement to obtain the independent isotopic proofs of the correct age determination makes the dating of sedimentary rocks so difficult.

Another aspect of the approach, suggested by Vinogradov, is as follows: "when speculating on the mixing lines, it is necessary to prove their actual existence and not just conceal behind them the inconvenient age parameters of the linear arrangement of data points" (Vinogradov, 1977; p. 102). In this instance, however, the aims of geochronological studies are misunderstood. In my opinion, what needs to be proved is the correctness of the defined age and its geologic meaning. Only if these confirmations cannot be obtained, we may consider other variants of interpretation of the straight-line dependence, the mixing line variant among them. But this is not the main point. The statement of Vinogradov that the mixing process is the only alternative to the age interpretation of linear arrangement of data points appears rather debatable, particularly if the meaning of the term "mixing" is clearly understood. According to Vinogradov, we may "find and imagine" interpretations of various kinds to explain the straight-line dependence incompatible with the age version. Possible variants are as follows: (1) proportional loss of radiogenic strontium in the course of superimposed processes; (2) the so-called "isochrones with initial inclination," which are quite likely in the case of samples of the fine-grained rocks; and (3) proportional mixing of ancient clastic and recent material newly formed by epigenetic processes, etc., including also the mechanism suggested by Gorokhov *et al.* (1995), for the straight-line arrangement of data points under conditions of a multicomponent system.

One more methodological aspect of the approach under discussion is the correspondence of the analyzed material to the concrete problem. It is evident that in order to date any geologic process it is necessary to study either the material formed as a result of the process, or the matter completely reworked by this process (in terms of the isotopic–geochronometric systems in use). From this point of view, clayey rocks are originally inappropriate for the determination of the time of their formation. The isotopic homogenization of strontium during the mechanical mixing of clastic particles with different Sr isotope composition and, consequently, with different Rb/Sr ratios should also result in averaging of these parameters. In this case, the isochron relations seem to be a sheer nonsense. If the Sr isotope homogenization depends on the interaction between clayey particles and cold sea water, then this homogenization will be more easy during any activation of the heated water solutions either present within the rock, or brought outside. If the readiness of this process corresponds to the natural situation, then the determination of the time of rock formation is likewise impossible.

The situation when the isotopic homogenization takes place in the course of sufficiently intensive epigenetic processes is more probable. Without dwelling on the principal problem of proving that exactly this event happened, let us discuss the possible consequences of the process and compare them with the available data. The complete Sr homogenization implies that all components of the rock at that moment were identical in their Sr isotope composition. On the other hand, since the selection of whole-rock samples is practically at random (at least, none of the papers by Vinogradov discuss the sampling criteria), the isochron model requirements suggest that Sr homogenization took place within the entire volume of the studied object. The first issue does not comply with data stating that any clayey rock includes both the more ancient particles (e.g., illites and feldspars), which retain the memory about the sourceland, and the younger components formed during the epigenetic processes, and this point is acknowledged by Vinogradov himself. Therefore, studying the clayey rocks, we always deal with a mixture of materials different in age and in $(^{87}Sr/^{86}Sr)_0$ values.

The complete homogeneity of Sr isotope composition in the entire rock sequence after the end of an epigenetic process can hardly be conceived, more so if the rocks remain sedimentary (during metamorphism, starting with the greenschist facies, the homogenezation of $(^{87}Sr/^{86}Sr)_0$ ratios is, apparently, actual). First, the separate parts of the sequence have different Rb/Sr relations and Rb and Sr concentrations, whereas, as shown by geochronological practice, the newly formed minerals, including illites and feldspars, often inherit the Sr isotope composition of the altered material. Second, the entire sequence cannot have the same permeability for water solutions and, consequently, an equally effective Sr isotope exchange between the rocks and fluid. Third, different places in the sequence may have different concentrations of the inherent clastic and newly formed materials.

Therefore, the straight line in the isochron coordinates would mean that all the components different in age in the randomly collected samples should be in a certain proportion. The inclination angle of this straight line can hardly correspond to the time of the real geologic event, even if the calculated age value seems quite reasonable. That is why an "isochron" obtained requires particular attention and certain proofs of a consistency between the calculated age and the time of the actual geologic process. By the way, it is not so difficult to obtain these proofs. It is sufficient to divide a pair of whole-rock samples into several components and analyze them. If the data points corresponding to these components plot on the isochron for the whole-rock samples, then they will prove the reality of the calculated age, the interpretation of which, however, requires

the additional mineralogical and stratigraphic information available.

This is not just an abstract reasoning as one can see from the results discussed in one of the papers by Vinogradov *et al.* (1994). In the course of dating sedimentary rocks from the Ognevskoe Group of eastern Siberia, glauconites and samples of sandstones, siltstones, and claystones were studied. The Rb–Sr isochron for glauconites corresponds to the age of 1275 ± 50 Ma. Therefore, the conclusion that the age of the host rocks is not younger than 1275 Ma is quite correct. The age determination for the Ognevskoe Group on the basis of whole-rock samples resulted as follows: the fine-grained rocks (six samples of siltstones and claystones) plot on the isochron, the inclination of which corresponds to 1526 ± 115 Ma. The data points of quartz–feldspar fractions from sandstones plot somewhat above the "isochron" for the fine-grained rocks, namely, along an errorchrone about 1600 Ma old. The authors suggest that 1526 Ma years ago the rocks experienced a stage of intensive transformations that completely homogenized the Sr isotope composition in siltstones and claystones; the transformations were insufficient, however, to homgenize the $(^{87}Sr/^{86}Sr)_0$ ratios in the coarse-grained rocks. In this case, the calculated age was attributed to the formation time of the new generation of K-feldspars. Consequently, the time of sedimentation was assumed to be older than 1526 Ma.

At first glance, this interpretation seems rather convincing. At least, the authors of the paper consider it likely that the straight line obtained for the fine-grained rocks is the mixing line. This assumption, however, is rejected because the considered data imply a lack of mixing components "with fixed Rb/Sr and $^{87}Sr/^{86}Sr$ values." This argument is not valid, since the description shows that the clastic material is composed of quartz and K-feldspars. It can be easily assumed that at the time of rock formation, the Rb/Sr ratios and the Sr isotope composition in the different parts of sedimentary rock were controlled only by the variations of these parameters in K-feldspars. Consequently, we should expect "an isochron with initial inclination." Therefore, in order to interpret the inference of 1526 Ma age as the possible maximum for the formation time of the Ognevskoe Group, it is sufficient to suppose that the K-feldspar, newly formed after the clastic material, inherited the geochemical and isotope characteristics of the latter. Of course, in the actual situation, a complete inheritance can hardly be expected, but the obtained straight line was not an isochron either. In order to convert it into the isochron, the authors had to reduce the accuracy of the Rb/Sr ratio determination by 1.5 times.

Skipping a few other considerations in favor of this version of interpretation, let us note that, according to it, the age of the studied rocks is invariably less than 1526 Ma and, what is more likely, it approaches the age of glauconites (1275 Ma). In addition, there seems to be no contradicting the stratigraphic data. Nonetheless, the two versions of interpretation have a common principal feature: they lack proof of the validity of the assumption on which they are based. In the version proposed by Vinogradov and his colleagues, it is supposed that the homogenization of the Sr isotope composition occurred in the course of the superimposed process; in my version, the isochron with initial inclination is proposed. The factual data that are at the disposal of the authors (and of the readers) do not provide grounds to prefer any of the two discussed, or other versions, which can be "found or imagined."

The Dating of Clayey Rocks Using Illite Fractions

The approach advocated by Gorokhov and Mel'nikov is more effective and methodologically consistent. First of all, it envisages a correspondence between the analyzed material and the problem at hand, because, in order to date a process, the material that was formed during this process should be selected. They chose illite, the properties of which as a geochronometer are quite adequate and well studied, while the applied methods offer a hope for successful identification of different generations of the mineral. Nonetheless, this approach is liable to criticism since it has certain disadvantages. I shall mention the most important.

(1) The results, obtained by Gorokhov and his colleagues, clearly show that the measured age of illites directly depends on the size of their fine fractions. As soon as one of the subfractions is identified as monomineral in composition (e.g., 0.1–0.2 μm), its measured age is assumed to be the time of the corresponding process. The size range of the subfraction, however, is arbitrary (e.g., <0.1; 0.1–0.2; 0.2–0.3 μm). Therefore, if the size ranges of fine fractions were assumed with a displacement (e.g., 0.05–0.15 and 0.15–0.25 μm) and one of them were identified as monomineral, then their measured age should correspond either to the average age of subfractions of <0.1 and 0.1–0.2 μm, or to the average age of subfractions of 0.1–0.2 and 0.2–0.3 μm. If the size ranges of subfractions are fixed arbitrarily, then we can hardly expect that one illite generation corresponds exactly to the given size interval. Therefore, the rock age should be limited to a certain range, and, lacking the knowledge about the proportions of different illite generations in the given subfractions, it is logical to restrict the range by corresponding mean values. In the case of the Lontova Formation dating (Gorokhov *et al.*, 1994), it would have been more correct, I believe, to date the early diagenesis not by the fixed value of 533 ± 8 Ma for the 0.1–0.2 μm subfraction, but within the range of 523–553 Ma restricted by the mean values for the corresponding subfractions. In this case, a more accurate age determination can be obtained only in the course of a more detailed differentiation of subfractions using those in which only a single illite generation is suspected. By the way, if it is correct to assume the monogeneration of illite, then, in the course of diminishing the subfraction sizes, the age–dimension plot

should show a gradient reflecting the real age of the illite monogeneration.

(2) Another principal limitation of the method is the indefinite time interval between the early diagenesis and sedimentation. This information is essential for using the available age data either for the exact correlation of rocks, or for calibration of the scale boundaries. The solution of this problem can be achieved by studying Cenozoic rocks with precisely established geologic ages, since the Cenozoic geological time scale is determined with sufficient accuracy and correctness, while the difference between the processes under consideration is comparable with the rock ages and unshaded by determination errors. For example, the paper by Ohr *et al.*, (1991) implies that the indicated time interval is about 15 m.y. long. The results of the first and now classical work on this problem, which was dedicated to dating of the Devonian black shales in Texas (Morton, 1985), show that a period of about 60 m.y. separates the sedimentation and illite formation stages. Obviously, uncertainties of this kind impose restrictions on solution of many problems in stratigraphy.

Apart from the restrictions already discussed, there are other problems requiring elucidation. One of them is related to the general dependence of nearly all crystallochemical parameters of illite on the subfraction size range. In this case, the monomineral state of one of the subfractions can be determined by using the electron microscopy. All other characteristics of illite indicate only the relation of the subfraction to the process responsible for its formation. Meanwhile, it is difficult to surmise that the subsequent stages of rock alteration developed, for example, under the decreasing temperature. However, this is the main factor responsible for the polytypes and crystallinity index of illite, and also for the dimensions of the formed individual minerals. General considerations indicate that the existence of different illite generations should be most obviously manifested in the form of steps and inflections in the plots mentioned above. Other problems concerning the determination of the amount of illite generations and their identification require the particular consideration.

Isotopic-Geochronological Methods in Stratigraphy

Summing up the discussion, I have to agree with both parties and to admit that, at present, it is practically impossible to determine an unambiguous isotope age for the accumulation time of clayey rocks. Unfortunately, these rocks cannot be used as yet for improvement of the geological time scale. This does not mean, however, that there is no sense in applying the isotopic–geochronological methods for stratigraphic purposes. Let us consider the possibilities of these methods for solving basic stratigraphic problems, such as correlation and elucidation of chronostratigraphic position of stratified deposits. Obviously, these problems can be solved, if the sequence under study includes rocks appropriate for the precise and unequivocal determination of their formation time. At present, only volcanic and volcano-sedimentary rocks (volcanic ashes, benthonites) are suitable for successful dating, as is evident from many papers, including those dedicated to improvement of the geological time scale (Williams *et al.*, 1982; Tucker *et al.*, 1990). These rocks should be the prime targets in stratigraphic research. Leaving aside the dating procedure for volcanigenic rocks discussed in detail in many publications, I would like to note some peculiarities of using of the results in geochronological studies.

If sedimentary sequences contain the indicated rocks, then their detailed correlation depends on the frequency of their beds in the section and on the analytical errors of age determinations. Under optimal conditions, the contemporary isotopic–geochronological methods yield the accuracy of a few to dozens of million years for Precambrian rocks, several million years for Paleozoic and Mesozoic rocks, and $n \times 10^{-1}$ million years for Cenozoic rocks. Naturally, these ranges also determine the correlation accuracy of volcano-sedimentary sequences.

While the problems of correlation can be solved by comparing the measured (or proven) age values, the position of the studied object in the standard stratigraphic scale can be determined by using the geological time scale, in which the boundaries are calibrated in absolute time units. Obviously, the quality of the scale controls the correctness of chronostratigraphic inferences, the adequateness of legends to geological maps, etc.. The state-of-the-art isotopic–geochronological substantiation of the Phanerozoic time scale is described by Rublev (1996) and that of the Precambrian scale by Semikhatov (1993). It is important to note that age values for many stratigraphic boundaries in the available Phanerozoic scales, including the most recent ones (Odin, 1994), require the additional substantiation and are, apparently, not always correct.

As far as it concerns the sedimentary rocks proper, I should mention some differences in applying the isotopic methods to Phanerozoic and Precambrian sediments. For example, the latter are considerably older and, consequently, display the relatively greater variations in isotopic parameters and essentially widen the capability of isotopic methods. For instance, the application of U–Pb and Pb–Pb methods to Precambrian carbonates is now feasible, though in this case we can determine only the time of the early diagenesis (Ovchinnikova *et al.*, 1995). On the other hand, the difference of more than 5–10 m.y. between the sedimentation time and the first stage of illite generation is too large to use the age of the latter in the solution of stratigraphic problems, but in the case of Precambrian rocks this difference is not as decisive, even if it is up to 50 m.y.

The potential of isotopic–geocronological methods in dating the sedimentary rocks proper still appear to be rather limited. The main reserve here is, perhaps, in the

further development of the approach suggested by Gorokhov and his colleagues. In addition, the geochronological potential of glauconites is not exhausted as yet, though we need reliable criteria for the diagnosis of glauconite fractions unaffected by secondary processes (Yakovleva and Gorokhov, 1996). Recently, there was a successful attempt to use the Sm–Nd method for dating the phosphate sediments (Yang Jie-Dong *et al.*, 1996). There is a hope that this approach will add an impulse to the solution of stratigraphic problems by means of isotope geochronology.

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